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Sm-Nd Isotopic Systematics of Metabasites from the Archean Sutam Complex of Granulites (Southern Aldan Shield)

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Abstract—The paper presents first data on the Sm-Nd isotopic systematics of metabasites from the supracrustal Sutam granulite complex (South Aldan Shield). It is established that the formation time of protoliths of mafic and ultramafic crystalline schists, which are conformable to the host enderbites, is 3067 ± 130 Ma. This date is close to the known U-Pb age of 3131 ± 74 Ma characterizing protolith of enderbite plagiogneisses of the Sutam complex. The result obtained suggests that the commencement of the early sialic crust formation in the southern Aldan Shield was not in the Early Archean, as had been customary accepted before, but near the Early–Late Archean boundary. The event was marked by an intense andesite-dacite volcanism and associated eruptions of basic and ultrabasic magmas.

Key words: *Sm-Nd isotopic data, metabasites, enderbites, Archean, Aldan shield*

INTRODUCTION

According to recent data (Drugova *et al.*, 1990; Nutman *et al.*, 1990; Sal'nikova, *et al.*, 1992; Neymark *et al.*, 1993, Kotov *et al.*, 1995), the deep basement of the Aldan Shield is heterogeneous and consists of infracrustal and supracrustal rock complexes variable in age. Four stages of the basement formation in the Aldan Shield correspond to the following time ranges: 3.6–3.3, 3.3–3.1, 2.8–2.6, and 2.4–2.0 Ma. The oldest rocks are distinguished in western and southwestern areas of the shield. Granulite complexes of the southern Aldan Shield, those of the Sutam block included, have been traditionally considered as Early Archean in age, but without any reliable support by isotopic dating. Recently, Shemyakin *et al.* (1998) measured the U-Pb age of enderbites from the Sutam block and obtained the date of 3131 ± 74 Ma. According to petrological study (Marakushev, 1965), granulites in the south of the Aldan Shield (Sutam, Zverev, Kurulta, and Dzhugdzhur blocks) belong to the deepest levels of the Earth's crust and are of particular interest for solving problems of the early evolutionary stages of the Aldan Shield.

In this work, our main objective was to study the Sm-Nd isotopic system of crystalline schists that are basic to ultrabasic in composition, and whose tabular bodies conformably rest inside the enderbite complex of the Sutam block.

GEOLOGY

The Sutam block located within the synonymous river basin is a part of the southern Aldan Shield. It is bounded by the Atugei-Nuyam graben in the north and separated from the Stanovoi foldbelt by the South-Aldan deep-seated fault in the south. In the west, its boundary corresponds to the Davangra-Khugda graben, and the Idzhek-Nuyam deep-seated fault represents the eastern limitation of the block (Fig. 1).

The Sutam block includes a series of second-order blocks locally displaying gentle thrust faults, which complicate the inner structure of the latter. Correlation of constituent metamorphic layers is difficult due to complex tectonic patterns of the main block.

Problems of lithological-stratigraphic subdivision of the Sutam metamorphic complex have been considered in a series of works (Kastrykina, 1974; Zedgenizov, 1974; Duk *et al.*, 1986; Mironyuk *et al.*, 1986, Nikitin and Akhmetov, 1990); however, the suggested stratigraphic schemes are poorly compatible. In addition, some researchers (e.g., Duk *et al.*, 1986) class the peculiar hypersthene plagiogneisses of the Sutam block with the infracrustal nonstratified rocks. On the basis of published data and our own observations, we arrived at the conclusion that the Sutam metamorphic complex is divisible into two principal units:

(1) the lower thick and monotonous subcomplex mostly consisting of hypersthene plagiogneisses (enderbites) with sporadic interlayers of metabasites;

(2) the upper sequence of hypersthene and other gneisses alternating with quartzites.

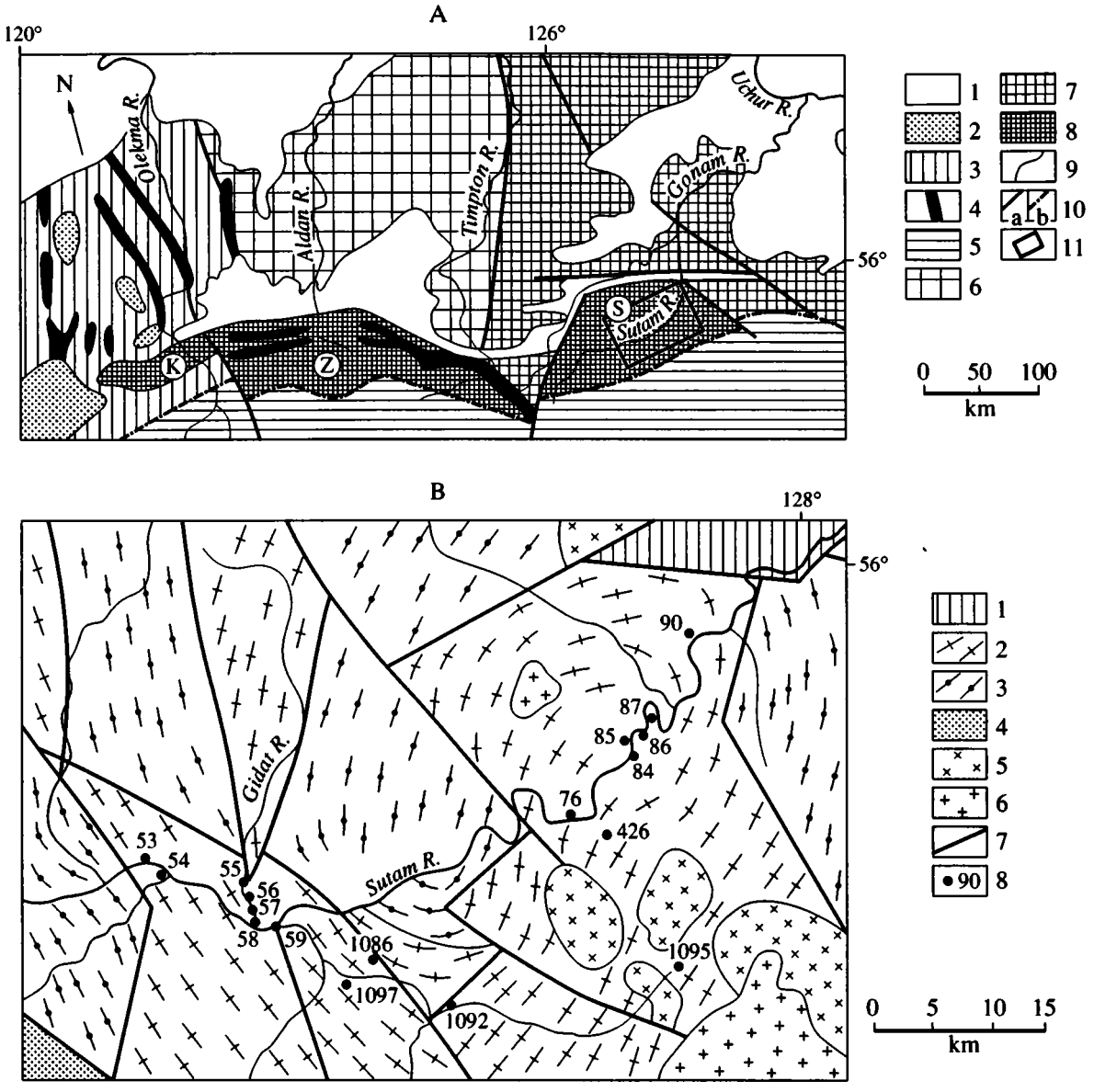

Fig. 1. Tectonic scheme of the southern Aldan Shield (A) and sampling sites in the Sutam River basin (B).

A: (1) sedimentary cover of the platform; (2) deposits in platform depressions of the aulacogen type; (3) Olekma granite-greenstone zone; (4) greenstone belts; (5) Stanovoi foldbelt; (6) West Aldan granulite-gneiss area; (7) East Aldan granulite-gneiss area; (8) Kurulta (K), Zverev (Z), and Sutam (S) granulite blocks of the southern Aldan Shield; (9) geological boundaries; (10) South Aldan (a) and other (b) faults; (11) study area. The scheme is compiled using and modifying the tectonic map of BAM area (Krasnyi, 1980).

B: (1) metamorphic rocks of the East Aldan granulite-gneiss area; (2) granulites of the lower Sutam subcomplex (enderbites and metabasites); (3) granulites of the upper Sutam subcomplex (High-Al gneisses, enderbites, quartzites, and metabasites); (4) sedimentary platform cover; (5) Proterozoic granitoids; (6) Phanerozoic granites; (7) faults; (8) sampling sites.

Our scheme of stratigraphic subdivision of the Sutam complex is closest to that suggested by Nikitin and Akhmetov (1990), who divided the complex into the concordant Nel'gyui and Seim formations. In terms of lithology and petrographic composition, the Nel'gyui and Seim formations are almost identical to the lower and upper subcomplexes in our scheme, respectively. The boundary between the latter is placed at the level, where interlayers of garnet to garnet-biotite plagiogneisses and magnetite quartzite show their first occurrence amid enderbites.

The lower subcomplex mostly consisting of enderbites has an apparent thickness of not less than 2000 m. Enderbites enclose individual interlayers (1–20 m thick), lentils, and boudins of metabasites and metaultrabasites whose total abundance in the sequence is 10–15%. Enderbites usually show fine- to medium-grained structures and are gray to dark gray in color, with a

greenish tint. Banding of these rocks is emphasized by thin melanocratic laminae rich in hypersthene crystals and locally associated hornblende. In places, enderbites are coarse-grained and less distinctly banded. The whole enderbite sequence is, however, well layered.

The upper subcomplex is variable in lithological composition. High-Al gneisses (of the garnet, garnet-biotite, garnet-sillimanite-cordierite, hypersthene-sillimanite, and other types) and enderbites are dominant rocks of the unit, and their layers are a few dozen to few hundred meters thick. Less abundant rocks are two-pyroxene plagiogneisses, various quartzites, and representatives of the eulysite rock group locally enclosing the iron ore deposits. Marble, calciphyre, and calcareous-siliceous schists occur sporadically. The subcomplex is about 2000 m thick.

Metamorphic grades typical of the Sutam rock complex are characterized in a series of publications (Marakushev, 1965; Kastrykina, 1974; Korikovskii and Kislyakova, 1975; Perchuk *et al.*, 1983; Duk *et al.*, 1986; Aranovich, 1991; Avchenko *et al.*, 1994). Rocks of the complex represent the deepest and high-temperature metamorphic facies, the so-called Sutam grade in terms of Marakushev (1965). Available data on *P–T* parameters of mineral paragenesis in metapelites of the Sutam complex reveal three thermal steps of their formation (Aranovich, 1991). For the early high-temperature stage, the estimated parameters are $T = 830–860°C$ and $P = 8–9$ kbar. This culmination of metamorphism was responsible for the origin of hypersthene-sillimanite-quartz paragenesis. The next step corresponds to $T = 680–760°C$ and $P = 4–5$ kbar. It resulted in the development of reaction structures widespread in metapelites and triggered processes of granitization. The third thermal step took place under $T = 540–580°C$ and $P = 3$ kbar. This one corresponds to the late stages of regressive metamorphism. In the present work, we investigated granulites from the lower Sutam subcomplex.

Hypersthene plagiogneisses (enderbites) most widespread in the lower subcomplex consist of plagioclase (35–50%; $An_{50–65}$), quartz (5–25%), hypersthene (3–8%), clinopyroxene (0–6%), and orthoclase (0–5%). Their accessory minerals are apatite, magnetite, and zircon.

Potassium feldspar and plagioclase usually show perthite and antiperthite intergrowth, respectively. Enderbites have been affected by K-metasomatism (charnockitization) to a considerable degree, and these rocks enclose the late generation of potassium feldspar (microcline) and sporadic biotite flakes. Amoeboid microcline grains locally constitute reaction rims around plagioclase relics of the early generation, which show the vermicular quartz intergrowths. Abundance of microcline varies from rare grains to 7–8% of the rock modal composition. According to petrographic observations, the acidic enderbite varieties ($SiO_2 = 65$ wt % or more) always contain the late microcline, though in variable proportions.

Two-pyroxene schists consist of plagioclase ($An_{50–58}$; 30–50% of the modal composition), clinopyroxene (15–40%), and orthopyroxene (5–10%). Some of the two-pyroxene plagiogneisses include accessory garnet or biotite. Other accessory minerals are apatite, zircon, titanomagnetite, and ilmenite.

Crystalline schists of ultramafic composition are composed of orthopyroxene (6–65% of the modal composition) clinopyroxene (9–25%), olivine (0–5%), and amphibole (0–60%). Ilmenite and biotite are their accessory minerals.

It is established on the basis of the distribution of major and trace elements (Mishkin and Lavrik, 1997; Mishkin *et al.*, in press) that volcanic rocks of the calc-alkali and komatiite-tholeiitic series represented protoliths of the lower Sutam subcomplex. Calc-alkali volcanics of basalt-andesite-dacite-rhyodacite series composed about 90% of the total rock volume.

Peridotitic komatiite and komatiitic to tholeiitic basalts are recognizable amid metavolcanic rocks of the komatiite-tholeiitic series.

ANALYTICAL METHODS

The following spectrum of analytical procedures was used to investigate the chemical composition and geochemistry of rock samples at the Far-East Institute of Geology: classical chemical method for major elements (analyst Z.S. Natarova); atomic absorption and quantitative spectral analyses for Cr and Ni (analysts G.A. Bakhareva, L.I Alekseeva, and L.I. Azarova); and neutron-activation technique for REE. Concentrations of Rb, Sr, Ba, Y, Zr, and Nb were measured at the Analytical Center of the Geological Institute, Russian Academy of Sciences (under the guidance of S.M. Lyapunov), using the fluorescence analytical procedure. Detection limits are as follows: less than 0.05 ppm for REE; 1 ppm for Rb, Sr, Nb, Zr, and Y; and 5 ppm for Ba. The relative analytical errors are 3% for Sr and REE, 4% for Ba and Y, 8% for Zr; 15% for Rb and Nb, 5–10% for Cr and Ni (atomic absorption method), or ~30% (quantitative spectral analysis in the case of low Cr and Ni content).

The Sm-Nd isotopic analysis was performed at the Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences, using the technique described earlier (Karpenko *et al.*, 1984).

The Nd isotope composition was measured using the modified "Cameca" mass spectrometer TSN-206 SA and applying the three-band (Re, W) ion source regime. Ion currents were measured for atomic masses 142, 143, 144, 146, and 150 by the Nd isotope analysis, and for atomic masses 147, 149, 152, and 154 by the Sm isotope analysis. In the first case, the possible presence of trace Ce (atomic mass 142) and Sm (atomic mass 149) ions was persistently controlled, and, when

Table 1. Chemical composition of granulites from the Sutam block

Major and trace elements	1097-i	1086-i	1086-l	1097-zh	57-b	87-b	86-a	90-b	84-a
	1	2	3	4	5	6	7	8	9
SiO_2	44.44	45.05	46.49	48.67	49	49.92	50.58	51.04	51.28
TiO_2	0.37	0.22	2.03	1.84	1.87	0.46	0.44	0.71	0.86
Al_2O_3	6.72	6.4	13.22	12.92	15.3	16.31	5.64	14.73	15.73
Fe_2O_3	4.87	5.68	5.82	9.98	2.23	0.89	2.15	n. a.	0.2
FeO	7.62	7.06	11.14	7.7	14.16	8.09	8.52	*12.66	9.83
MnO	0.32	0.23	0.26	0.23	0.24	0.18	0.24	0.24	0.17
MgO	30.47	29.99	7.88	6.54	4.2	9.54	23.78	7.56	7.62
CaO	4.64	3.87	10.06	9.4	9.35	11.36	6.49	9.41	10.27
Na_2O	0.42	0.41	2.54	2.26	2.8	2.39	0.84	2.94	3.32
K_2O	0.01	0.1	0.32	0.2	0.41	0.19	0.39	0.71	0.49
P_2O_5	0.12	0.09	0.24	0.26	0.24	0.07	0.08	n. a.	0.13
L. O. I.	n. a.	0.9	n. a.	n. a.	0.2	0.6	0.85	n. a.	0.2
Cr	3300	4100	44	23	80	600	3500	252	270
Ni	890	1200	55	50	35	200	1000	109	250
Rb	<1	6.3	6.7	3	<1	4.3	5.6	5	9.2
Ba	43	67	160	150	130	72	99	355	345
Sr	24	40	98	130	125	79	54	150	210
Nb	<1	1	6.8	6	5.5	1.8	1.2	5.5	2.6
Zr	30	22	90	100	93	26	33	58	47
Y	12	5.7	30	28	31	14	10	28	17
La	3	4	8.1	7.8	10	8.9	12	35	9.8
Ce	6.4	7.5	19	19	23	15	22	72	18
Nd	4.5	3	14	14	16	8.9	9.4	42	11
Sm	1.1	0.68	4.2	3.9	4.5	1.7	1.7	7.1	2.8
Eu	0.29	0.17	1.4	1.2	1.5	0.5	0.46	1.2	0.86
Tb	0.25	0.12	0.96	0.87	1	0.4	0.31	0.98	0.5
Yb	1	0.51	3	2.7	3.6	1.8	0.9	3	2.2
Lu	0.16	0.087	0.45	0.43	0.51	0.27	0.12	0.41	0.34

Note: Oxides in wt %; trace element concentrations in ppm; * total iron content as FeO; n.a., not analyzed. Columns: (1,2) olivine-two-pyroxene-amphibole schists; (3, 4, 9) two-pyroxene schists; (5) garnet-two-pyroxene schist; (6–8) two-pyroxene-amphibole schists. Rocks are analyzed at the Far East Institute of Geology (analysts Z.S. Natarova, G.A. Bakhareva, L.I. Alekseeva, and L.I. Azarova) and at the Analytical Center of Geological Institute, Russian Academy of Sciences (under guidance of S.M. Lyapunov).

necessary, the analytical data on Nd isotope composition were respectively corrected. The measured Nd isotopic ratios were normalized relative to $^{150}Nd/^{142}Nd$ = 0.209627, and value of $^{149}Sm/^{154}Sm$ = 0.60750 was used to normalize the measurement results for the Sm isotope composition. The $^{147}Sm/^{144}Nd$ ratio is defined at the precision level of 0.3%.

RESULTS AND DISCUSSION

Samples for Sm-Nd isotopic study have been collected from tabular bodies of metaultrabasites (olivine-two-pyroxene crystalline schists) and metabasites (two-pyroxene crystalline schists), which concordantly rest inside enderbites of the lower Sutam subcomplex (Table 1).

Samples weighing about two kilograms each included rock specimens selected from different layers of the lower subcomplex using the "nest" sampling procedure in order to avoid the granitization effect. Sampling sites are shown in Fig. 1b.

The results of Sm-Nd isotopic study obtained for the whole rock samples of granulites from the Sutam block are presented in Table 2 and plotted on the Sm-Nd evolutionary diagram (Fig. 2). All the data points plot along an isochron that corresponds to the age of 3067 ± 130 Ma and mark the eruption time of calc-alkali and

Table 2. Sm-Nd isotopic data for granulites of the Sutam Complex

Sample no.	Sm, ppm	Nd, ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd(+-2\sigma)$
M-57-b	4.20	13.60	0.1869	0.51174 +-4
M-84-a	1.68	6.83	0.1490	0.51101 +-5
M-87-b	1.90	9.28	0.1249	0.51054 +-6
M-90-b	6.37	34.70	0.1073	0.51016 +-5
M-86-a	1.34	6.58	0.1236	0.51050 +-5
R-1097-zh	4.16	13.10	0.1917	0.51193 +-4
R-1097-i	1.45	5.25	0.1663	0.51134 +-5
R-1086-l	4.02	13.30	0.1829	0.51171 +-5
R-1086-i	0.67	3.07	0.1352	0.51078 +-6

komatiite-tholeiitic lavas. The isochron is characterized by positive value of $\varepsilon_{Nd}(0) = +2.7$ (MSWD = 1.93).

Geological data show that the original intermediate to acidic calc-alkali volcanics and basalts represented an integral lithological-structural complex subsequently affected by metamorphism. Consequently, there is no reason to expect a considerable age difference between the rocks. The Rb-Sr isotopic date of 2856 ± 140 Ma points to the time of granulite metamorphism that affected volcanogenic protoliths of the lower Sutam subcomplex (Mishkin *et al.*, in press).

The Sm-Nd date obtained for the Sutam metabasites is close, within the analytical error, to the U-Pb age of 3131 ± 74 Ma measured for enderbites of the complex (Shemyakin *et al.*, 1998). Thus, our results and published data suggest that the commencement of the early sialic crust formation in the southern Aldan Shield was not in the Early Archean, as had been customary accepted before, but near the Early–Late Archean boundary, as the latter is defined in the General stratigraphic scale of the Precambrian authorized by the Second All-Union conference of 1990 in Ufa (Semikhatov *et al.*, 1991).

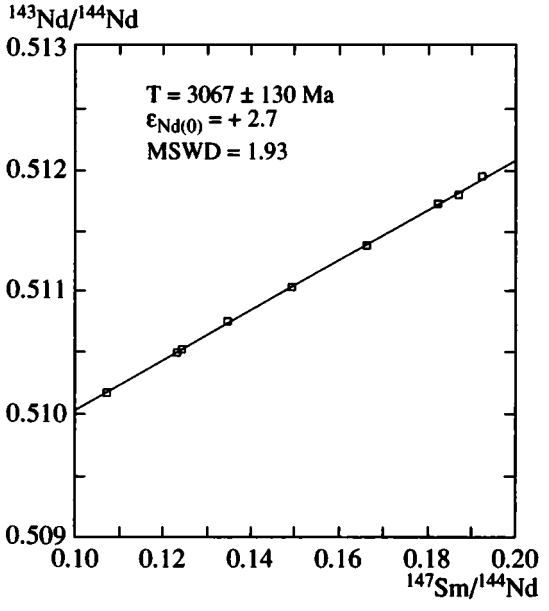

Fig. 2. Sm-Nd evolutionary diagram for granulites of the Sutam Complex.

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Stratigraphy of Paleozoic Shaly-Siliceous and Volcanosedimentary Deposits and Geologic History of the Polar Urals

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Abstract—Conodonts found in shaly-siliceous and volcanosedimentary deposits of different tectonic zones of the Polar Urals provided a new insight into the Paleozoic history of the region. Shaly-siliceous deposits of the Central subzone of the Lemva zone (the Chernaya Gora Formation), which were attributed earlier to the Middle–Late Ordovician, are of the Middle Ordovician–Late Devonian age. The Grubeshor (the Middle Ordovician–Silurian) and Malyudshor (the Devonian) formations are distinguished in volcanogenic-siliceous deposits of the Grubeshor subzone. The Orang Formation of the Manitanyrd-Paipudyna zone, which is usually considered as Ordovician in age, should be attributed on the basis of conodont findings to the Famennian–Carboniferous. New data on the Paleozoic stratigraphy of the Polar Urals stimulated us to revise the idea that the region is a characteristic example of the passive continental margin. The Polar Urals represent a structure that developed in three destructive stages: the Late Cambrian–Tremadocian, Middle–Late Ordovician, and Late Devonian. It is inferred that a new oceanic basin, which is proposed to be named the Uralian–Arctic, emerged in the Late Devonian–Carboniferous at the junction of the East-European continent and the Middle Paleozoic accretionary system of the Urals.

Key words: *stratigraphy, Paleozoic, conodonts, tectonic zone, Polar Urals*

Different structural complexes are tectonically imbricated in the region of the Polar Urals. The complexes form a nappe system, where sections are different in facies (Voinovskii-Kriger, 1945; Dembovskii, 1981; Shishkin, 1987; Ruzhentsev and Saveliev, 1997). The lower element of the system is the carbonate rock group (the relative autochthon, O–C) that is overthrust by allochthonous slices of mainly terrigenous and siliceous deposits (ϵ_3–C), and also of magmatic complexes (including ophiolites), which represent crust fragments of the Uralian paleoocean (Saveliev and Samygin, 1979; *Formirovanie...*, 1986; Ruzhentsev and Didenko, 1998). Clearly, the paleogeographic and geodynamic reconstructions in such a complicated structure are possible only on the basis of the detailed stratigraphy of different, structurally isolated facies complexes, especially of shaly to cherty and volcanosedimentary deposits widespread in the Polar Urals. Their stratigraphic subdivision and correlation of distinguished horizons were carried out on the basis of discovered conodont faunas. Conodonts are rare in the Uralian Lower Paleozoic deposits, and their species composition is not diverse. The Ordovician and Silurian conodonts from the Polar Urals are characteristic of the Baltic-Scandinavia region and Canada. Conodonts in the Devonian deposits are abundant, diverse and represented by cosmopolitan taxons characterizing zonal assemblages of the standard conodont scale. The results of our studies are presented in this paper.

For the sake of convenience in presenting the actual material, we briefly outline the tectonic zones of the Polar Urals (Fig. 1), which are termed as follows (from west eastwards): the Elets zone (carbonate rock group series, O_1–C); Lemva zone (siliceous-terrigenous deposits, ϵ_3–C); Manitanyrd-Paipudyna zone (the outlier of Precambrian basement of the Proto-Urals with a sedimentary cover of terrigenous, carbonate, and siliceous rocks, O_1–C); Kharbei zone (metamorphic rocks of the basement), Nauntin-Nyrdvomenshor zone (ophiolites and volcanosedimentary deposits, D_3–C); Shchuchiya-Voikar zone (ophiolites of the Syum-Keu, Rai-Iz and Voikar-Synya massifs and volcanosedimentary deposits, O–D). Our studies were carried out in the Lemva, Manitanyrd-Paipudyna and Nauntin-Nyrdvomenshor zones. The type sections of these zones are characterized below.

STRATIGRAPHY OF SHALY-SILICEOUS AND VOLCANOSEDIMENTARY SEQUENCES

The Lemva zone. Four subzones—Western, Central, Grubeshor and Eastern or Nelka (Fig. 2)—are distinguished within this zone (Ruzhentsev and Saveliev, 1997). The terrigenous sequence (*Pogurei Formation*) variable in thickness, composition, and rock grain size

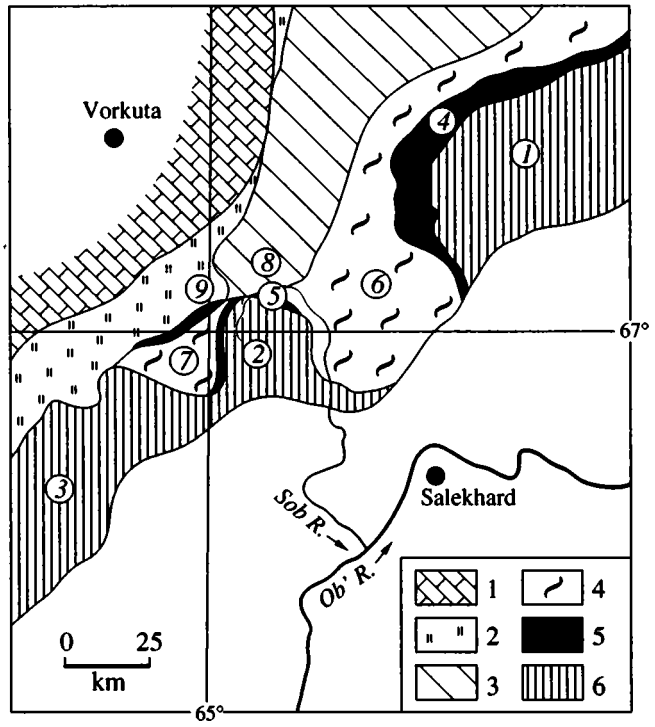

Fig. 1. Schematic tectonic zoning in the Polar Urals (at the latitude of the Sob transverse uplift). Tectonic zones: (1) Elets; (2) Lemva; (3) Manitanyrd-Paipudyna; (4) Kharbei; (5) Nauntin-Nyrdvomenshor; (6) Shchuchiya-Voikar. Numbers in circles: (*1*) Syum-Keu, (*2*) Rai-Iz, and (*3*) Voikar-Synya ophiolite massifs; (*4*) Nauntin-Pe Mountains; (*5*) Nyrdvomenshor melange; (*6*) Kharbei and (*7*) Kharamatalou metamorphic complexes; (*8*) Bol'shaya Paipudyna River basin; (*9*) Goluboi Creek basin.

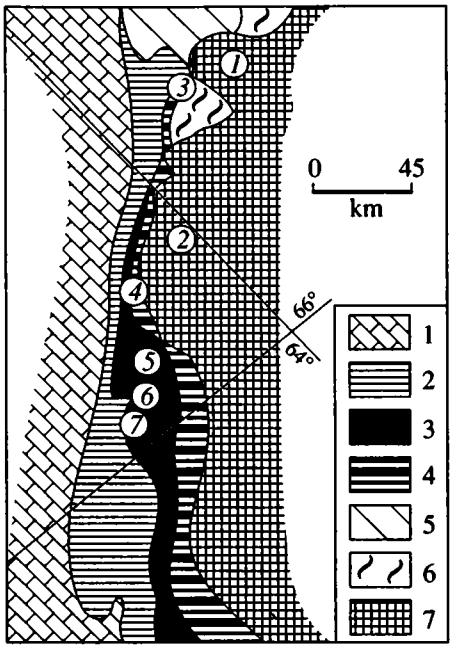

Fig. 2. Schematic tectonic zoning in the southern part of the Polar Urals. Tectonic zones: (1) Elets; (2) Western; (3) Central, (4) Grubeshor and Eastern subzones of the Lemva zone; (5) Manitanyrd-Paipudyna; (6) Kharbei; (7) ophiolites. Numbers in circles: (*1*) Rai-Iz and (*2*) Voikar-Synya ophiolite massifs; (*3*) Golubaya River basin; (*4*) Grubeshor River basin; (*5*) Paga River basin; (*6*) Pogurei River basin; (*7*) Samson Mountains.

is basal one in each subzone. In places (e.g., in the Central subzone), terrigenous deposits associate with rocks of the bimodal basalt-rhyolite volcanogenic series (*Kokpela Formation*). Until recently, the age of the Pogurei Formation was determined as Tremadocian on the basis of numerous finding of brachiopods (Voinovskii-Kriger, 1945; Dembovskii *et al.*, 1990). In the Pogurei River basin, we have found the Late Cambrian conodonts in calcareous sandstone beds corresponding to the lower part of the formation (Fig. 3: Site 1 with *Proacontiodus jiliensis* (Chen et Gong), *Cordylodus proavus* Mull, and *Eoconodontus notchpeakensis* (Mill.) of the Late Cambrian *proavus* Zone preceding the Ordovician; sites 2 and 3 with *Phakelodus* cf. *tenuis* (Mull.); determinations by S.V. Dubinina). Accordingly, we attribute the host deposits to the Late Cambrian–Tremadocian. The formation is 0.3–2 km thick.

The Grubeyu Formation (O_1a) is exposed in three western subzones and consists of variegated phyllitic aleuropelites (up to 600 m thick) alternating with tuffites. The Eastern subzone accumulated at that time a sequence (up to 1.5 km thick) of variegated phyllites and differentiated volcanics (from basaltic andesites to rhyolites), in which lavas are intercalated with numerous horizons of tuffs, volcanomictic sandstones and conglomerates. Another characteristic feature of this zone was the accumulation of siliceous deposits during the Late Ordovician, Silurian and Devonian.

The Western subzone. The presented description of the Western subzone is brief, since its characteristics were specially discussed in a series of works (Koren', 1972; Eliseev, 1973; Puchkov, 1979; Shishkin, 1987; Saldin, 1996; and others). Exposed here from the base upward are the following rocks: black clayey to clayey-siliceous shales, cherts and limestones of the Kachamylk Formation (O_{2-3}); clayey-siliceous shales, cherts and limestones of the Kharota Formation ($S–D_1$); sandstones, limestones, and cherts of the Paga Formation (D_{1-3}); clayey-siliceous shales, cherts, and limestones of the Nyanvorga Formation ($D_3–C_1$); clayey shales and limestones of the Yaiyu Formation (C_{1-2}); and flysch of the Kechpel Formation ($C_3–P_1$). The section yielded various paleontological remnants concentrated in carbonate deposits.

The Central subzone is characterized by a wide distribution of carbonate-free, carbonaceous, clayey-siliceous and siliceous rocks distinguished as the Chernaya Gora Group. The group is correlative with the

Chigim Formation supposed to be the Middle–Late Ordovician in age (Gesse *et al.*, 1976). The data obtained in the course of our studies allow us to revise the stratigraphy and age of these deposits. In terms of lithology, these uniform terrigenous-siliceous rocks form a series of tectonic slices, and it is difficult to restore their original sequence. The deposits are best exposed in the upper courses of the Pogurei River within the Chernaya Mountain area (at the mount with the altitude of 707 m). The Middle Ordovician–Famennian conodonts of several tectonic slices were found in separated sites shown in Figs. 2 and 3 (Ruzhentsev *et al.*, 1996). Judging from the conodont ages, three groups of slices can be distinguished within the subzone: the Middle Ordovician–Early Silurian, Silurian–Early Devonian, and Middle–Late Devonian.

Slices of the first group (O_2–S_1) are exposed at the Chernaya Mountain summit and in foots of its eastern slope. Black clayey-siliceous to clayey shales intercalated with phthanite beds (up to 1 m thick) are exposed here. The rocks contain conodonts (Fig. 3) *Drepanoistodus subrectus* (Br. et Mehl) of the Middle Ordovician at the Site 4; *Oulodus* sp. of the Late Ordovician–Early Silurian at the Site 5; *Pseudobelodella silurica* Armst., *Belodella* cf. *cooperi* Armst., *Carniodus* cf. *carnulus* (Wallis.), *Belodella* cf. *silurica* Barric, *Distomodus staurognathoides* Wallis., *Pterospathodus* sp., and *Ozarkodina* sp., (S_1ln_3-w_1) at the sites 6 and 7; and *Pterospathodus pennatus* (Wallis.), *Ozarkodina* sp., *Belodella* sp., and *Panderodus unicostatus* Br. et Mehl., (S_1ln_3-w_1) at the Site (8) The apparent thickness of the deposits does not exceed 350 m.

The Silurian–Lower Devonian deposits are exposed along the eastern slope of the Chernaya Mountain, where their section consists of the following units (from the base upward):

(1) Clayey-siliceous shales with thin (5–10 cm) phthanite interlayers and conodonts *Ozarkodina* sp. and, *Panderodus* sp. (S–D_1), Site 9. The apparent thickness is up to 80 m.

(2) Platy, plicate and cataclastic phthanites up to 20 m thick.

(3) Alternation of phthanite interlayers (up to 10 cm thick), black when fresh and light gray on the weathered surface) with medium-grained quartzose sandstones; conodonts detected in phthanites at the Site 11 are *Pandorinellina optima* (Mosk.), *Ozarkodina remscheidensis* (Ziegel.), and *Panderodus* sp. (D_1l). The thickness is about 130 m.

(4) Gray quartz sandstones, medium-grained and coarsely bedded. The apparent thickness is up to 70–75 m.

A similar section is exposed in the watershed of the Pogurei and Pogureiegat rivers, where it consists of alternating clayey-siliceous shales, cherts, and mudstones up to 150–170 m thick. Their sequence is likely overturned; lower cherty beds (Site 12) yielded conodonts *Polygnathus foliformis* Snig. and *Icriodus* sp.

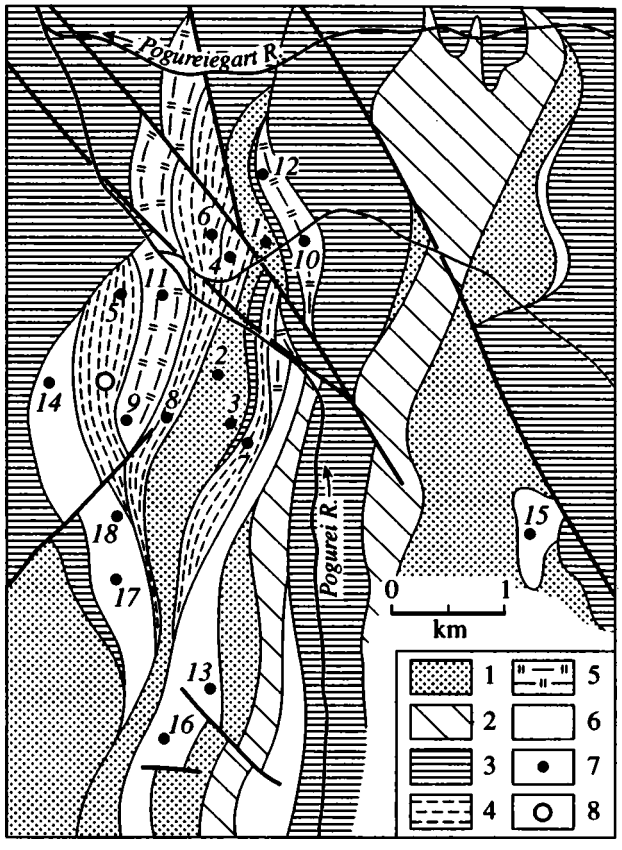

Fig. 3. Geological Sketch map of the Pogurei River basin: (1) Pogurei Formation, $\in_3$–O_1t; (2) Kokpela Formation, $\in_3$ (?)–O_1t; (3) Grubeyu Formation, O_1a; (4–6) Chernaya Gora Group (O_2–D_3fm) and its subdivisions corresponding to (4) O_2–S_1, (5) S–D_1, and (6) D_{2-3}; (7) conodont sampling sites (see text for numbers); (8) Chernaya Mountain (height mark 707 m).

(D_1e), whereas those from the upper (Site 10) part are identified as *Pandorinella optima* (Mosk.), *Ozarkodina remscheidensis* cf. *repetitor* Ziegel., and *Belodella* sp. (D_1l-p).

The Middle–Upper Devonian deposits are recognized in upper reaches of the Pogurei River, where the rock sequence exposed on the left slope of the river valley is as follows (from the base upward):

(1) Carbonaceous-siliceous shales and black mudstones. The apparent thickness is 13 m.

(2) Black medium-grained and coarsely bedded quartz sandstones. The thickness is 30 m.

(3) Alternating clayey and clayey-siliceous shales and phthanites bearing conodonts (Site 13) *Polygnathus* ex gr. *linguiformis* Hinde, *P. trigonicus* Bisch. et Ziegel., *P. angustipennatus* Bisch et Ziegel., *Tortodus* sp., *Panderodus* sp. (D_2e). The thickness is up to 60 m.

(4) Quartz sandstone 26 m thick.

(5) Clayey-siliceous to clayey shales with interlayers of recrystallized gray cherts (Site 16); conodonts of

the site are classed with *Polygnathus* sp. (D_{2-3}). The thickness is up to 90–100 m.

It is likely that rocks exposed along the right bank of the Pogurei River represent the middle part of the section. The member of interlaminated clayey-siliceous shales and phthanites is exposed here (Site 15) is about 15–17 m thick and yields conodont species *Polygnathus* ex gr. *varcus* Stauf. (D_2zv). The member underlies clayey-siliceous shales with interlayers of black quartz sandstones; the last rock sequence is up to 120 m thick.

The upper part of the section under discussion was studied in the western slope of the Chernaya Mountain, where the Arenigian variegated phyllites are overridden with the tectonic contact by the following rocks:

(1) Clayey-siliceous shales with numerous interlayers (up to 10 cm thick) of gray cherts; conodonts of the unit (Site 14) are *Panderodus* sp. and *Polygnathus* sp. (D_2). The apparent thickness is 8–10 m.

(2) Black mudstones 25–27 m thick.

(3) Alternating black quartzose sandstones and phthanites bearing conodonts (Site 17) *Ancyrognatus* sp. and *Polygnatus* sp. (D_3f). The thickness is 25 m.

(4) Black, coarsely bedded quartz sandstones 40–45 m thick.

(5) Gray laminated quartz sandstones with "floating" fragments (up to 1 cm in size) of phthanites and clayey-siliceous shales. The thickness is 3–4 m.

(6) Gray laminated cherts with intercalations of yellowish-gray and black clayey-siliceous shales; conodonts (Site 18) *Icriodus* sp. and *Polygnathus perplexus* Thomas (D_3fm_2) are found here in association with redeposited Middle Devonian and Frasnian forms. The apparent thickness is 16–18 m.

The Famennian deposits are best represented several kilometers southwestwards, in the Samson Mountains (Fig. 2, Site 7), where they form a core of a large antiform. The rock sequence exposed within the latter is as follows (from the base upward):

(1) Uniform black phthanites and clayey-siliceous shales enclosing beds (up to 3 m thick) of gray quartz sandstones; conodonts identified in cherts are *Polygnathus* sp., and *Palmatolepis* sp. (D_3). The thickness is 180 m.

(2) Alternation of gray to light gray radiolarites, black phthanites, gray clayey-siliceous shales with intercalations (up to 0.5–1 cm) of green-gray, gray and black mudstones; conodonts *Palmatolepis glabra prima* Ziegl., *P. glabra lepta* Ziegl. et Huddle, *P. minuta loba* Helms (D_3fm_1, the Upper *crepida* Zone) are found in upper cherty interbeds. The thickness is 130–140 m.

(3) At the higher structural level (the contact is unexposed), there are thin-bedded black phthanites, gray cherts, black to yellowish-gray and almost white in places (bleached) siliceous shales; conodonts *Palmatolepis rugosa ampla* Mull. (D_3fm_2, the *postera-expansa* zones) are found 60–70 m above the sequence base; the early Famennian redeposited conodonts are also present there. The thickness is up to 250 m.

Similar facies of Devonian deposits are widespread in the watershed of the Paga and Palniktyvis rivers (Fig. 2, Site 5). Terrigenous-siliceous deposits are highly dislocated here and form a series of tectonic slices. On the right slope of the valley, in the Paga River upper reaches, basalts of the Kokpela Formation (ϵ_3-O_1t) and variegated phyllites (O_1a) are overlapped by the following rocks:

(1) Platy cherts and clayey-siliceous shales with conodonts *Palmatolepsis triangularis* Sann., *P. minuta loba* Helms, *P. delicatula* Br. et Mehl, and *Icriodus alternatus* Br. et Mehl (D_3fm_1); in addition, the Emsian-Frasnian redeposited conodonts are also found in the rocks. The apparent thickness is about 45 m.

(2) Bedded plicated phthanites intercalated with thin interlayers of black clayey shales; conodonts *Ozarkodina remscheidensis remscheidensis* (Ziegl.) are recognized in the deposits (D_1). The apparent thickness is 20–25 m.

(3) Clayey-siliceous shales with interlayers of cataclastic phthanites bearing conodonts *Ozarkodina* sp., *Pandorinellina* sp., *Icriodus* sp., and *Belodella* sp. (D_1). The thickness is 110–115 m.

(4) Mudstones and black siliceous mudstones with interlayers (0.5–1 m) of quartz sandstones in the upper part of the unit. The thickness is 180–190 m.

(5) Quartz sandstones 15–18 m thick.

(6) Alternation of quartz sandstones and clayey-siliceous shales. The thickness is 60–65 m.

(7) Phthanites and clayey-siliceous shales with conodonts *Polygnatus* cf. *varcus* Stauf. (D_2zv_2). The thickness is 8–10 m.

(8) Quartz sandstones. The apparent thickness is 17–20 m.

Despite the fragmented state of the studied sections, it is possible to show that the time range of the Chernaya Gora Group is from the Middle Ordovician to Famennian. All series and most stages distinguished in this time interval are substantiated by the composition of the discovered conodont assemblages. The composite section of the group is very uniform with regard to lithology. The condensed clayey-siliceous sediments dominate in the group. Quartz sandstones probably representing deposits of grain flows appear in the Silurian to be widely developed in the Devonian. The presence of redeposited conodont forms, together with the absence of horizons affected by rewashing, are indicators of deep-water hiatuses manifested to a maximum extent in the Late Devonian.

The Grubeshor subzone is an area of widespread Ordovician–Devonian volcanogenic-siliceous deposits, which make up a series of tectonic slices thrusted over rocks of the Central subzone. The Ordovician–Silurian part of the section is distinguished as the Grubeshor Formation, and the Devonian interval corresponds to

the Malyudshor Formation (Gesse *et al.*, 1976). This subdivision scheme, however, appears to be inadequately justified, because the section is quite uniform, poor in fauna, and complicated in structure.

We studied the most complete section of the *Grubeshor Formation* in the northern part of the subzone along the Goluboi Creek (Fig. 2, Site 3). The structure of the section is as follows (from the base upward):

(1) Variegated phyllites similar in lithology to those of the Grubeyu Formation (O_1a); the apparent thickness is 200–250 m.

(2) Alternation of violet phyllites, black clayey- and clayey-siliceous shales; the Middle Ordovician conodonts occur in clayey-siliceous shales (personal communication of P.M. Kucherina). The thickness is 18–20 m.

(3) Clayey-siliceous shales and thin-platy siliceous siltstones. The thickness is 80–90 m.

(4) Schistose, dark to greenish gray spillites enclosing members (up to 3–5 m thick) of black clayey-siliceous shales and local phthanite beds (0.5–0.7 m); conodonts *Panderodus liratus* Nowl. et Barn. (O_3-S_1) are recognized in the middle part of the unit. The apparent thickness is 300–350 m.

A similar section, though devoid of fauna, was observed in the southern part of the subzone, in the Vostochnaya Kokpela, Igyadeiegard and Vostochnaya Tambulava river basins.

The Malyudshor Formation is of spillites alternating with carbonaceous-siliceous shales, phthanites, and more rare quartz siltstones, among which members of lithoclastic tuffs are present in places (the Srednyaya Lagorta, Paga and Malyudshor river basins). Limestone lenses comprising the Late Silurian–Early Devonian crinoid fragments were detected in some localities (the Srednyaya Lagorta River basin). Conodonts *Pandorinellina steinhornensis* (Ziegl.), *P.* ex gr. *optima* (Mosk.), and *Polygnathus* cf. *inversus* Klap. et Johns (D_1e) were identified in cherts intercalated with spillites in the section located in the Srednyaya Lagorta River valley.

An interlayer of siliceous shales in similar deposits in the upper courses of the Paga River yielded conodonts *Mesotaxis asymmetricus* (Bisch. et Ziegl.), *Polygnathus webbi* Stauf., *P.* ex gr. *decorosus* Stauf., and *Belodella* sp. (D_3f_1). The Malyudshor Formation is hundreds of meters thick.

Tectonic slices of uniform cataclastic gray to light gray bedded cherts frequently occur among basalts of the Grubeshor subzone. As a rule, they are free of interlayers of volcanic rocks. Taking into account the structural position, we include them arbitrarily into the section of the Grubeshor subzone. The cherts are dated back to the Famennian. For instance, conodonts *Palmatolepis rugosa ampla* Mull. and *P. gracilis sigmoidalis* Ziegl. (D_3fm_2, the *postera* Zone), as well as the Early Devonian, Givetian, Frasnian, early Famennian redeposited elements, are macerated from these cherts in the right slope of the Grubeshor River valley.

According to the aforesaid, the stratigraphic range of volcanogenic-siliceous sequence of the subzone corresponds to the Middle Ordovician–Frasnian interval, being similar, in general, to that of the Chernaya Gora Group. We don't include the Famennian deposits into the Malyudshor Formation, because they represent probably an independent stratigraphic unit.

The Eastern (Nelka) subzone is poorly studied. The time interval under consideration is represented here by a sequence (up to 800 m thick) of barren quartzose to oligomictic sandstones and siltstones (the Nelka Formation). The sequence mainly includes beds of frequently calcareous rocks with well-graded clastic material. The rocks continue with a gradual transition to the Arenigian deposits. For this reason, we arbitrarily attribute the lower part of the sequence to the Middle Ordovician. The upper time limit of the Nelka sandstones is not established.

In order to conclude the description of the Lemva zone, we should point out that the Middle Ordovician–Famennian deep-water sequences of different facies accumulated within its three western subzones. The sedimentation took place in the basin with a deep axial part (carbonate-free clayey-siliceous and volcanogenic-siliceous sequences) and relatively shallow-water marginal areas (substantially carbonate sediments of the Western and Eastern subzones).

The Manitanyrd-Paipudyna zone. Basement rocks of the Proto-Urals and thick (up to 150 m) basal conglomerates are overlaid with unconformity by carbonaceous-terrigenous (O) and carbonate (S–D_2) sequences exposed in vast areas within the western part of the zone (Dembrovskii *et al.*, 1990; Ruzhentsev and Aristov, 1998, and unpublished data). All these deposits are of shallow-water origin and bear benthonic fauna. The deposits are similar to coeval strata in the Elets zone. The character of the rock sequence changes in the Late Devonian.

We studied the Frasnian section in the watershed of the Razvilnyi and Diyavol'skii creeks (right tributaries of the Bol'shaya Paipudyna River, Fig. 4, Site 1), where it includes the following units (from the base upward):

(1) Green or violet basaltic flows intercalated with members of epiclastic rocks. The thickness is 120 m.

(2) Gray medium-grained quartz-calcareous sandstones enclosing marble blocks with the Early–Middle Devonian conodonts. The thickness is 25–30 m.

(3) Arenaceous sandstones with conodonts *Icriodus symmetricus* Br. et Mehl, *Palmatolepis* sp., and *Polygnathus* sp. characteristic of the Frasnian Stage of the Upper Devonian. The thickness is 12 m.

(4) Rhythmic alternation of quartz sandstones, siltstones, and pelitomorphic limestones; the unit yielded Late Devonian, most likely, the Frasnian conodonts

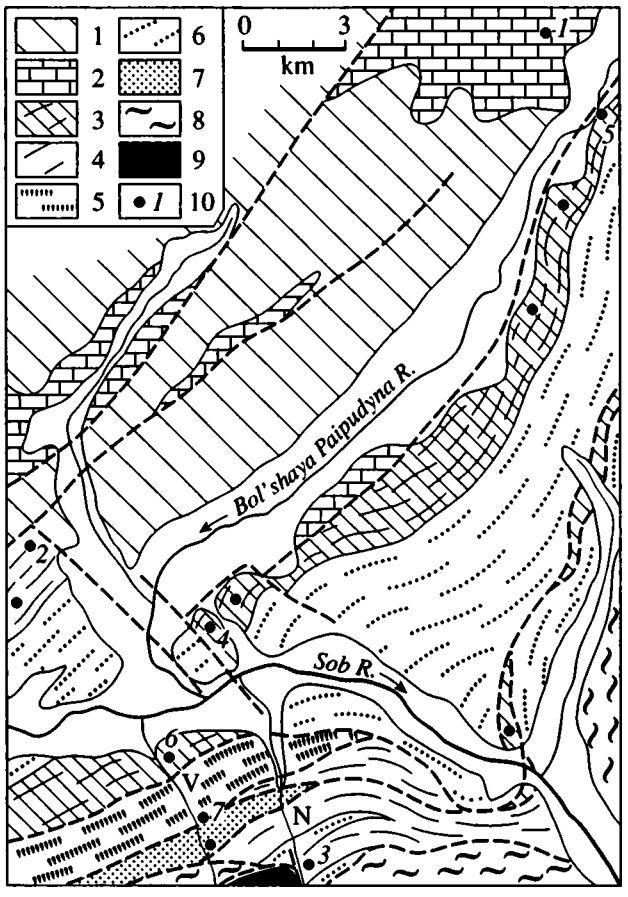

Fig. 4. Geologic scheme of the Sob and Paipudyna river basins: (1) Proto-Uralian basement (Pt_3), (2) terrigenous-carbonate series (O_1–D_3f), (3) Nyanvorga (D_3fm–C_1) and Yaiyu (C_{1-3}) formations, (4) lower (D_3fm–C_1t) and (5) middle (C_{1-3}) subdivisions of the Orang Formation, (6) variegated (D_3) and siliceous-shaly (D_3fm–C_1) sequences of the Voishor Formation, and (7) graywackes of the Rai-Iz Formation (C_1v) in the Manitanyrd-Paipudyna zone; (8) Kharbei zone; (9) Nauntin-Nyrdvomenshor zone; (10) sections described in the text; (V) Voishor Creek; (N) Nyrdvomenshor Creek.

Polygnathus sp., *Palmatolepsis* sp. The thickness is 170 m.

(5) Alternation of calcareous sandstones, clayey-siliceous shales, and more rare phthanites. The thickness is 70–75 m.

The appearance of carbonate-terrigenous turbidites and siliceous rocks in the upper part of the section indicates a transition from neritic to bathyal conditions of sedimentation. The bathyal sequence is of the Famennian–Carboniferous age. Deposits of this age are characteristic of the eastern part of the zone. They are intensely dislocated and form a system of slices thus impeding compilation of the composite section. The Famennian–Carboniferous deposits are represented by three facies: siliceous-shaly (the Orang Formation), carbonaceous-siliceous-shaly (the Nyanvorga and Yaiyu formations), and terrigenous (the Voishor and Rai-Iz formations).

The Orang Formation consists of three siliceous, siliceous-shaly and sandy-siltstone units. A band of their exposures in the Paipudyna River basin extends along the right bank of the Sob River headstream (north of the Rai-Iz Massif). We studied the section at the junction of Bol'shaya and Malaya Paipudyna rivers (height marks 381.0 and 336.0 m, Fig. 4, Site 2), where the following units are exposed:

(1) Alternation of platy gray cherts, phthanites, clayey-siliceous shales, and siliceous siltstones; conodonts of the unit are *Neopolygnathus communis* (Br. et Mehl), *Palmatolepsis minuta minuta* Br. et Mehl, *P. gracilis gracilis* Br. et Mehl., and *P. gracilis sigmoidalis* Ziegl. (D_3fm_{2-3}). The apparent thickness is 70 m.

(2) Phthanites alternating with clayey-siliceous and clayey shales; abundance of clayey shale beds increases upsection. Conodonts *Idiognathodus* sp. and *Gondolella* sp. (C_{2-3}) were recognized in the phthanite interlayer (5 cm thick) occurring in the middle part of the unit; in addition to these species, the interlayer yielded redeposited conodonts *Palmatolepsis minuta minuta* Br. et Mehl, *P. gracilis gracilis* Br. et Mehl, *Neopolygnathus communis* (Br. et Mehl), *Siphonodella* sp., *Pseudopolygnathus* sp. (D_3–C_1t), *Lochriea* cf. *commutata* (Br. et Mehl), and *L. multinodosa* (Wirth) (C_1v-s). The apparent thickness is 230–250 m.

(3) Alternation of clayey-siliceous and clayey shales with quartzose and oligomictic sandstones. The thickness is 120–130 m.

(4) Oligomictic medium-grained, locally calcareous sandstones alternating rhythmically with quartz siltstones. The apparent thickness is 500–600 m.

A similar section is exposed in the right bank of the Sob River (the middle course of its right tributary Nyrdvomenshor Creek; Fig. 4; Site 3):

(1) Coarse-bedded brecciated phthanites bearing poorly preserved conodonts *Palmatolepsis* ex gr. *minuta* Br. et Mehl and *P.* sp. (D_3fm). The apparent thickness is 10–12 m.

(2) Thin-platy ("varved") gray to black, highly plicated cherts with conodonts (D_3fm_{2-3}) *Palmatolepsis gracilis gracilis* Br. et Mehl, *P. gracilis sigmoidalis* Ziegl., *P. minuta minuta* Br. et Mehl, *Neopolygnathus communis* (Br. et Mehl). The thickness is 40–50 m.

(3) Platy cherts alternating with clayey-siliceous shales and bearing conodonts *Neopolygnathus communis* (Br. et Mehl) and *Siphonodella* cf. *praesulcata* Sand. (D_3fm_3-C_1t_1). The thickness is 40–45 m.

(4) Gray to greenish-gray phyllites enclosing interlayers of black clayey-siliceous shales and cherts, in places, lenses of quartz sandstones; Carboniferous conodonts of the unit are identified as *Gnathodus* sp. (*Paragnathodus* sp.?). The thickness is 250 m.

(5) Gray to greenish-gray quartz or oligomictic, sometimes micaceous sandstones with interlayers of

quartz siltstones and shales. The apparent thickness is up to 800 m.

The Orang Formation has long been regarded as Ordovician in age. Our data show that the cherty horizon is attributable to the Famennian, the siliceous-shaly beds to the Carboniferous, and the sandstone-siltstone strata to the Late Carboniferous. Taking into account a considerable thickness of the latter (more that 1 km), we assume the Permian age for the uppermost beds.

The Nyanvorga (D_3-C_1) and *Yaiyu* (C_{1-3}) formations constitute the carbonate-siliceous-shaly sequence of the zone. Deposits of this kind constitute large tectonic slices among rocks of the Orang Formation. The fragmentary state of the section is evident, and relationships between these formations could not be observed during the field work.

The Nyanvorga Formation section was studied in the watershed of the Sob and Bol'shaya Paipudyna rivers (height marks 286.6 and 168.0 m north of the railway bridge; Fig, 4, Site 4). Deposits are dislocated into a series of faulted folds, and the succession of strata is given below with certain reservations.

(1) Clayey-siliceous shales and phthanites; conodonts recognized in the deposits are *Palmatolepsis triangularis* Sam., *P. delicatula* Br. et Mehl, *Ancyrognathus* cf. *asymmetricus* (Ult. et Bassl.), and *Icriodus alternatus* Br. et Mehl (D_3fm_1). The apparent thickness is 30–32 m.

(2) Clayey-siliceous shales, phthanites, siliceous siltstones, and siliceous limestones (lenses up to 30 cm thick); conodonts *Palmatolepsis gracilis gracilis* Br. et Mehl, *P. gracilis sigmoidalis* Ziegl. (D_3fm_{2-3}) are identified in limestones. The thickness is 130–140 m.

(3) Clayey-siliceous shales with thin (up to 1–1.5 m) interlayers of quartz-feldspathic sandstones. The apparent thickness is 60 m.

A band of carbonaceous-siliceous-clayey deposits extends from the region mentioned above further northwards for 30 km along the left bank of the Bol'shaya Paipudyna River. The deposits are poorly exposed. Some fragments of the section were studied in scarps of the river valley. For instance, scarps of the left bank of the Bol'shaya Paipudyna River (opposite to the Mramornyi Creek mouth; see Fig. 4, Site 5) exhibit a dislocated member of platy phthanites, carbonaceous clayey-siliceous shales, siliceous siltstones, and siliceous limestones, whose apparent thickness is up to 80 m. Conodonts *Streptognathodus* sp., *Gnathodus* sp. (C_2), as well as redeposited conodonts *Palmatolepsis gracilis sigmoidalis* Ziegl. (D_3fm_{2-3}) were recognized in phthanites. The deposits in question correspond to the Yaiyu Formation, although they are identical in lithology to the Famennian (Nyanvorga) deposits.

We restored a more characteristic section of the Yaiyu Formation on the basis of rocks exposed along the Voishor Creek (the right tributary of the Sob River; see Fig. 4, Site 6). The units distinguished here in the valley scarps are characterized below (from the base upward):

(1) Clayey and clayey-siliceous shales. The apparent thickness is 10–12 m.

(2) Gray bituminous pelitomorphic limestones with shale interlayers; the lower limestone interlayer yielded conodonts *Gondolella* s. l. (C_2m-C_3) in association with redeposited *Pandorinellina optima* (Mosk.), *P. philipi* Klap. (D_1l-p), Polignathus sp. (D_1e–D_2), *Eognathodus* cf. *sulcatus* Phil. (D_1p); *Polygnathus* sp. (D_1e–D_2), and *Gnathodus bilineatus* (Roundy) (C_1v-s). The thickness is 7 m.

(3) Clayey-siliceous shales 28–30 m thick.

(4) Limestones similar to those of Bed (2). The thickness is 24 m.

(5) Shales with thin intercalations (1–2 cm) of siliceous limestones. The apparent thickness is 20 m.

The terrigenous (Voishor) type of the section is represented most completely in the middle and upper courses of the Voishor Creek (see Fig. 4, Site 7). Deposits are dislocated into a series of narrow folds complicated by numerous longitudinal ruptures, and relationships between some elements of the section are not clear. According to our data, the section includes three members.

The lower member:

(1) Greenish-gray, olive-colored, and violet phyllites and fine-grained sandstones with thin (up to 5 cm) interlayers of gray siliceous shales (tuffites?). The deposits resemble in appearance those of the Grubeyu Formation (O_1a) from the Lemva zone and are correlated with them. The apparent thickness is 150 m.

(2) Alternation of green, violet, and gray phyllites with black clayey and clayey-siliceous shales. The bed yielded conodonts *Palmatolepsis rugosa grossi* Ziegl. and *Polygnathus* sp. (D_3fm_2) in association with redeposited Frasnian elements. The thickness is 12–15 m.

The middle member:

(3) Clayey and clayey-siliceous shales with single interlayers (up to 10 cm thick) of phthanites. The thickness is 20 m.

(4) Clayey-siliceous to clayey shales and siltstones with interlayers of phthanites and siliceous limestones.

Further upstream, the deposits change only slightly. They are represented mainly by numerous wedges of intensely sheared clayey-siliceous shales, phthanites, siliceous siltstones, and limestones. The latter play a subordinate part as separate interlayers 2–10 cm thick. The middle member is similar to the Nyanvorga Formation in the rock types and their stratification.

The upper member (the Rai-Iz Formation) is exposed in the upper courses of the Voishor Creek, where graywacke turbidites with an apparent thickness of more than 500 m have a gradual transition to the carbonate-siliceous-shaly strata (D_3fm-C_1t). Fine aleuropelitic sediments occur at the base of the section and grade upward into sandy turbidites with lenses of grav-

elstones and conglomerates, the clastic material of which is derived from rocks of the Kharbei zone and ophiolites of the Rai-Iz Massif. The lower part of the member is attributed to the Visean.

To conclude characterization of deep-water deposits of the Manitanyrd-Paipudyna zone, we should point out that the spatial and structural relationships of the carbonate-free (Orang) and carbonate-siliceous (Nyanvorga) deposits of the Famennian–Carboniferous age are not very clear. The latter likely correspond to the outer (western in modern coordinates) shallow part of the basin. Therefore, the carbonate-free siliceous-shaly sequences accumulated in the inner deep-water zone of the condensed sedimentation. We should also emphasize a steady increase of terrigenous material in the Carboniferous sediments. Siliceous sediments give way at this level to siliceous-terrigenous and then to terrigenous deposits. Changes of this kind characterize different time intervals. For instance, the event occurred during the Early Carboniferous in the Voishor section (the graywacke flysch is of the Visean age there), but marks the boundary between the Middle and Late Carboniferous in the Nyanvorga (carbonate-siliceous) and Orang (siliceous-shaly) sections. The aforesaid allows a conclusion that deposits of the Voishor type accumulated in the easterly part of the zone, whereas graywackes of the Rai-Iz Formation mark the eastern edge of the deep-sea trough, where voluminous clastic material derived by erosion from the Kharbei and Nauntin-Nyrdvomenshor zones was deposited since the Early Carboniferous.

The Nauntin-Nyrdvormenshor zone. In the structural position, the zone corresponds to lower allochthons of the Syum-Keu and Rai-Iz ophiolite massifs. The same structures likely extend south of the Rai-Iz massif into the western flank of the Voikar-Synya Massif (the Yaiyu River basin).

The Sum-Key and Rai-Iz massifs represent synforms, each consisting of two packets of tectonic sheets. The section of upper sheets is more complete and includes all elements of the ophiolite association. Their basaltoid part dates back to the Late Silurian–Early Devonian (Afanasiev, 1990; Sharma *et al.*, 1995). In the lower sheets of tectonic melange, basaltoids are reckoned to be of the Ordovician–Early Silurian age. We studied the northwestern part of the Syum-Keu Massif (Nauntin-Pe Mountains), where the section of the lower allochthon includes the following rocks (from the base upward):

(1) Serpentinite melange with blocks of gabbroids, amphibolites, diabases.

(2) Green and violet schistose spillites and diabases with pillow-lave texture retained in places; abundant diabase bodies and impregnation zones of serpentinite were also observed. The apparent thickness is up to 500 m.

(3) Alternation of basalts and their epiclastic varieties; the rocks are schistose; the upper part of the section includes individual interlayers (up to 5 cm thick) of sheared marbles with conodonts (site at the height mark 231.8 m) of genera *Polygnathus* (of the Famennian appearance) or *Siphonodella* (D_3fm-C_1t). The thickness is 350–400 m.

(4) Alternation of quartz-sericite to quartz-graphite schists with sheared microquartzites (cherts), schistose tuffs and basalts. The apparent thickness is up to 200 m.

We attribute, with some reservations, Units 2 and 3 to the Late Devonian–Early Carboniferous and Unit 4 to the Carboniferous.

We obtained similar results studying the Rai-Iz Massif. The section compiled for the upper courses of the Nurdvomenshor Creek (the northern part of the massif) includes the following units (from the base upward):

(1) Cherts and clayey-siliceous shales of the Orang Formation (D_3fm-C_2).

(2) A horizon of serpentinite melange up to 300–350 m thick. The serpentinite matrix encloses blocks of the Orang Formation rocks, harzburgites, gabbro, gabbro-diabases, and basalts.

(3) A sequence of greenish gray aphyric basalts often represented by pillow lavas, aquabreccia, and, more rarely, by massive flows. Diabase bodies are abundant in the lower part of the sequence. Sedimentary interlayers are very rare and represented by phthanites and siliceous tuffites. One cherty lentil from the lower part yielded conodont species *Ozarkodina* cf. *sannemanni* Bish. et Ziegl. (presumably D_2zv_2-D_3f_1). In the middle part of the sequence (the southern slope of Skal'naya Mountain), we detected conodonts *Palmatolepsis transitans* Mull., *P. subrecta* Mull. et Youngq., *Ancyrodella* cf. *rotundiloba* (Bryant), *Mesotaxis assymetricus* (Bisch. et Ziegl.), *Polygnathus webbi* Stauf., and *P. decorosus* Stauf. (D_3f_1). The thickness is up to 1000 m.

(4) Greenish-gray tephroturbidites with rhythms as thick as 50 cm. The thickness is 60–65 m.

(5) A variegated horizon of alternating greenish-gray tephroturbidites and violet aleuropelites; also present are interlayers of basaltic tuffs and more rare flows of plagioporphyrites. The thickness is up to 90 m.

(6) Alternation of siliceous siltstones, clayey-siliceous to clayey shales, and phthanites enclosing single flows of aphyric basalts. Phthanites bear conodonts *Palmatolepsis rhomboidea* Sann., *P.* cf. *crepida* Sann., *P. subperlobata* Br. et Mehl, *P. minuta minuta* Br. et Mehl, *P. minuta loba* Helms, *P. gracilis gracilis* Br. et Mehl, *P.* cf. *subgracilis* Bisch., and *P. perlobata schindewolfi* Mull. (D_3fm_1). The apparent thickness is 70–80 m.

(7) Alternation of tephroturbidites, clayey-siliceous shales, siliceous siltstones, and limestones; conodonts from this level are *Palmatolepsis gracilis gracilis* Br. et Mehl, *P. gracilis sigmoidalis* Ziegl., and *P. minuta*

minuta Br. et Mehl (D_3fm_2). The thickness is 110–120 m.

Horizon 7 represents an upper element of the lower Rai-Iz allochthon. It is overthrust by ultramafic rocks of the upper allochthon being transformed into serpentinite melange. Basalts of Bed 3 are of the early Frasnian age; deposits of units 4 and 5 are likely attributable to the upper Frasnian, and those of beds 6 and 7 to the Famennian. The Syum-Keu and Rai-Iz sections show some similar and also distinctive features. Both are of the same structure: massive basalts occur at the base, whereas the volcanosedimentary units (tephroturbidites, epiclastic rocks, clayey-siliceous shales) rest at the top. It is evident the basalts are as old as the Late Devonian. The volcanosedimentary rocks from the Rai-Iz Massif are dated back to the Famennian, those from the Syum-Keu Massif correspond to the Famennian-Carboniferous interval. We cannot rule out that the mentioned age difference is related to incompleteness of paleontological records. However, an alternative inference is quite possible. As was noted above, the Rai-Iz Massif appeared at the erosion level as early as in the Early Carboniferous; as a result, a frontal zone of graywacke flysch (the Rai-Iz Formation) accumulated the erosion products derived from ophiolites. The same process began in the Syum-Keu sector only in the Late Carboniferous, and this was responsible for the considerably younger age of volcanosedimentary rocks in this area.

PECULIARITIES OF GEOLOGIC HISTORY OF THE POLAR URALS

The presented material on the stratigraphy of Paleozoic deposits in the Polar Urals allows a correlation of rock sequences in different zones of the region, providing a key for deciphering the geologic history of the Uralian paleoocean. The basin development started in the Late Cambrian, when, in response to rifting, a system of grabens filled with clastic material different in composition and size (the Pogurei Formation) was formed on the Proto-Uralian (Baikalian or Cadomian) basement. The close spatial and temporal connections between sequences of terrigenous rocks and bimodal basalt-rhyolite volcanites (the Kokpela Formation) supports this point of view. The axial graben that separated the East European continent from the Manitanyrd-Kharbei microcontinent (the Manitanyrd-Paipudyna and Kharbei zones) corresponds, in the modern structure, to the eastern subzone of the Lemva zone. The commencement of graben closure was in the Tremadocian, and as a result, a series of differentiated (from basaltic andesites to rhyolites) volcanites was formed in the Arenigian. We regard the latter as products of a suprasubduction volcanic belt originated along the eastern (in present-day coordinates) edge of the East European continent (Ruzhentsev and Didenko, 1998).

At the same time, the Lemva trough (three western subzones) originated west of the volcanic belt. The trough represented a vast basin filled with fine-grained silty-clayey and terrigenous sediments (the Grubeyu Formation), the composition and structure of which indicate a maximum degree of differentiation due to the remoteness of the provenance.

The geodynamic situation changed in the Middle Ordovician, when the East European continent collided with the Manitanyrd-Kharbei microcontinent. At that time, the East Lemva volcanic belt became extinct, and a sequence of sialic terrigenous deposits (the Nelka Formation) marking the eastern margin of the Lemva trough accumulated here. Since that time, the trough represented a deep intracontinental basin, within which the condensed clayey-siliceous deposits of the Chernaya Gora Group accumulated under conditions of an overall transgression. The composition of sediments points to their accumulation below the carbonate compensation depth (CCD). Shishkin (1987) calculated that the depth was close to 2.5 km. The Middle and Late Ordovician time and was specific in the history of the Urals. It was just at that time when oceanic basins of the Sakmara, Magnitogorsk, Denisovka, and Tagil zones originated. Development of the axial volcanic belt of the Lemva trough (the Grubeshor and Malyudshor formations) was a reflection of destructive processes in the Polar Urals. Unlike the South Urals, the Ordovician destruction in the belt did not give rise to the oceanic crust formation (corresponding ophiolites are missing here).

A new evolutionary stage of the Polar Urals commenced at the Middle–Late Devonian transition. The distinguished Upper Devonian–Carboniferous ophiolites (the Nauntin-Nyrdvomenshor zone) made it possible to suggest that the destructive processes migrated at that time eastward, where a new oceanic basin originated at the junction of the East European continent and adjacent Middle Paleozoic accretionary system of the Urals. It has been suggested to name the newly formed structure as the Uralian-Arctic ocean. Its origin was accompanied by formation of the passive continental margin. Mainly carbonate sediments were still accumulating at that time within the shelf platform (the Elets zone). Bathyal sediments (the Nyanvorga, Yaiyu, and Orang formations) accumulated in the Lemva and Manitanyrd-Paipudyna zones. In the Manitanyrd-Paipudyna zone, the replacement of carbonate sediments ($S–D_2$) by siliceous deposits ($D_3–C_3$) correlates well with the appearance of concurrent ophiolites. Since the Frasnian, the Lemva and Manitanyrd-Paipudyna zones can be regarded as elements of a complicated slope of the East European continent. The general cessation of tectonic activity in structures under discussion began in the Carboniferous in response to the formation of the Hercynian thrust-and-fold structures, whose development terminated in the Permian.

CONCLUSION

Conodonts found in shaly-siliceous and volcanosedimentary deposits of different tectonic zones of the

Polar Urals led to a new understanding of the Paleozoic stratigraphy of the region. It should be pointed out that conodonts are the only fossils in these sediments that could foster dating and subdivision of the sequences. Along with conodonts, radiolarians were found in cherts, but they are completely recrystallized, deformed and unsuitable for identification even to the generic level. Shaly-siliceous deposits (the Chernaya Gora Group) of the Central subzone of the Lemva zone, which were earlier attributed to the Middle–Upper Ordovician, now can be considered as the Middle Ordovician–Upper Devonian in age on the basis of conodonts findings. All series and most of stages of this stratigraphic interval are distinguished. The Grubeshor (Middle Ordovician–Silurian) and Malyudshor (Devonian) formations are recognized in volcanogenic-siliceous deposits of the Grubeshor subzone. The Orang Formation of the Manitanyrd-Paipudyna that was referred before to the Ordovician turned out to be yielding the Famennian–Carboniferous conodonts.

New data on the stratigraphy of Paleozoic sedimentary and volcanosedimentary deposits of the Polar Urals discredit the idea that the region is the example of the standard passive continental margin. The Polar Urals represents a structure, the Middle Paleozoic history of which developed in three destructive stages ($\in_3$–O_1t, O_{2-3}, D_3–C) and their time ranges are substantiated to a sufficient extent. Each stage corresponds to its own series of rock complexes, which accumulated parallel to the reorganization of structural patterns in the region. In the Carboniferous, loci of destructive processes presumably migrated eastward, where a new oceanic basin originated at the junction of the East European continent and adjacent Middle Paleozoic accretionary system of the Urals. We propose to name the newly formed basin as the Uralian-Arctic ocean.

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Stratigraphy of the Silurian Type Areas in East Siberia

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Abstract—Three stratigraphic type areas (the Moiero River vicinity, Turukhansk and Noril'sk districts), where clayey–carbonate sedimentary facies are widespread, characterize the Silurian of East Siberia. The Moiero River neighborhood is a type area of the Moierokan, Khaastyr, Agidy (representing together the Llandovery), and Khakoma (an equivalent of the Wenlock) horizons, whereas the Tukal (Gorstian Stage) and Postnichnyi horizons (Ludfordian–Pridoli interval) are distinguished in the Turukhansk and Noril'sk districts, respectively. The type areas exhibit stratotypes of all Silurian horizons, which are simultaneously the stratotypes of synonymous formations of East Siberia. The formations are subdivided into subformations, inside which the local chronozones are recognized. The chronozones are described on the basis of elementary geological bodies ("geothemes"), whose sequences are shown in the respective profiles. Structural and facies variations are exemplified for 6 formations, 13 subformations, and 54 local chronozones.

Key words: *Silurian, stratigraphy, type area, horizon, chronozone*

As a stratotype region, East Siberia includes the Siberian platform and Taimyr Peninsula. During the Silurian, the epicontinental sedimentary basin of the platform type occupied the whole region. Facies of a deep to shallow open shelf and of shoal–bar, lagoonal (closed shallow shelf), and coastal zones are distinguished amid sediments of the basin (Tesakov *et al.*, 1979). The terrigenous deep-shelf sediments are known in the northern Taimyr and in northwestern and central parts of the Tunguska syneclise. The terrigenous–carbonate facies of the open shallow shelf and shoal-bar zone are widespread in the central Taimyr and along limbs of the Tunguska syneclise. Characteristic of southwestern and southern areas of the Siberian platform are carbonate–halide and sandy facies of the closed shallow shelf. In accordance with sequences of local stratigraphic units (formations), the distribution zone of Silurian deposits in East Siberia is divided into 15 stratigraphic areas (Tesakov *et al.*, 1997, Fig. 1). Three of the latter (the Moiero River vicinity, Turukhansk and Noril'sk districts) represent the type areas, where the shallow–shelf facies are widespread. These facies rich in indicative and abundant remains of fauna and flora (Tesakov *et al.*, 1998a, 1999) are suitable for correlation with both the deep- and shallow-water deposits. Formations, subformation, and chronozones distinguished in the stratotype areas are described below. The description is based on data characterizing the elementary geological bodies or "geothemes," as they have been termed by Tesakov (1978).

The type area of the Moiero River and adjoining upper courses of the Kotui River are situated in the northeastern margin of the Tunguska syneclise (Fig. 1). This is the type area of the Moierokan, Khaastyr, and Agidy horizons spanning the time range of the Llandoverian Stage, and also of the Khakoma Horizon corresponding to the Wenlockian Stage. Stratotypes of the horizons are identical to those of synonymous formations. The Llandoverian and Wenlockian deposits of the area have been well characterized earlier, in works by Nikiforova and Andreeva (1961), Myagkova *et al.* (1963, 1977), and Tesakov *et al.* (1979, 1985). Brief characteristics of their formations and subformations are presented below, together with a description of local chronozones whose sequence within subformations has been observed in profiles exposed along the Moiero River (Fig. 2).

The **Moierokan Formation** (Fig. 3) has been distinguished in 1979 (Tesakov *et al.*, 1979) and termed after the Moierokan River, the right-hand tributary of the Moiero River. Its stratotype is 111.0 m thick and includes exposures 82 (beds 1–23) and 83 (beds 24–28). Studying the formation in 1979, we initially considered as typical the disconnected exposures N52–N69 and P74–P78, but later in 1985, the original stratotype description was supplemented with data on exposures P78–P79 (Tesakov *et al.*, 1985, p. 39, beds 9/5–9/7, and Fig. 15, beds 79/1–79/3), P78–P82 (Ibid., p. 41 and Fig. 15, beds 4–24), and P78–P83 (Ibid., p. 50 and Fig. 15, beds 25–28). Deposits of the formation are exposed in the middle courses of the Moiero River and penetrated by boreholes Chirinda-1 and Chirinda-2.

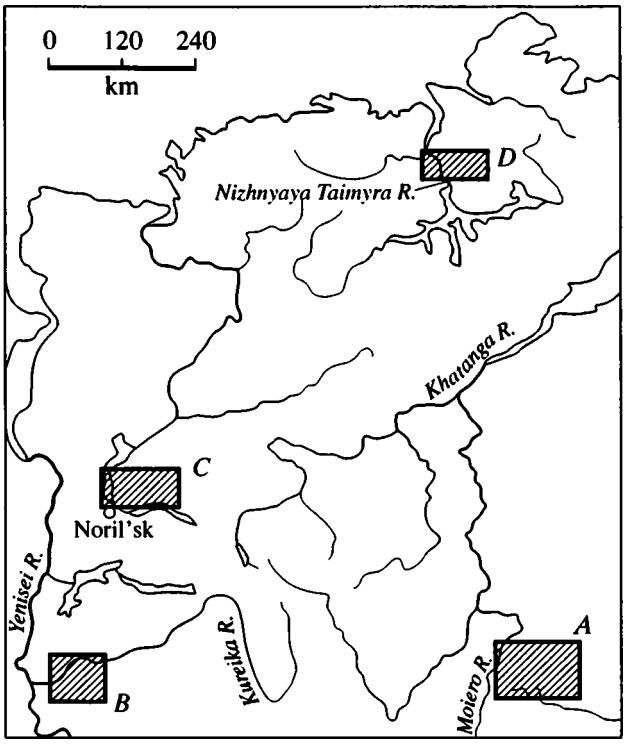

Fig. 1. Stratigraphic type areas of Silurian deposits in East Siberia: (*A*) Moierokan; (*B*) Turukhansk (Kureika River); (*C*) Noril'sk; (*D*) Middendorf (parastratotype section).

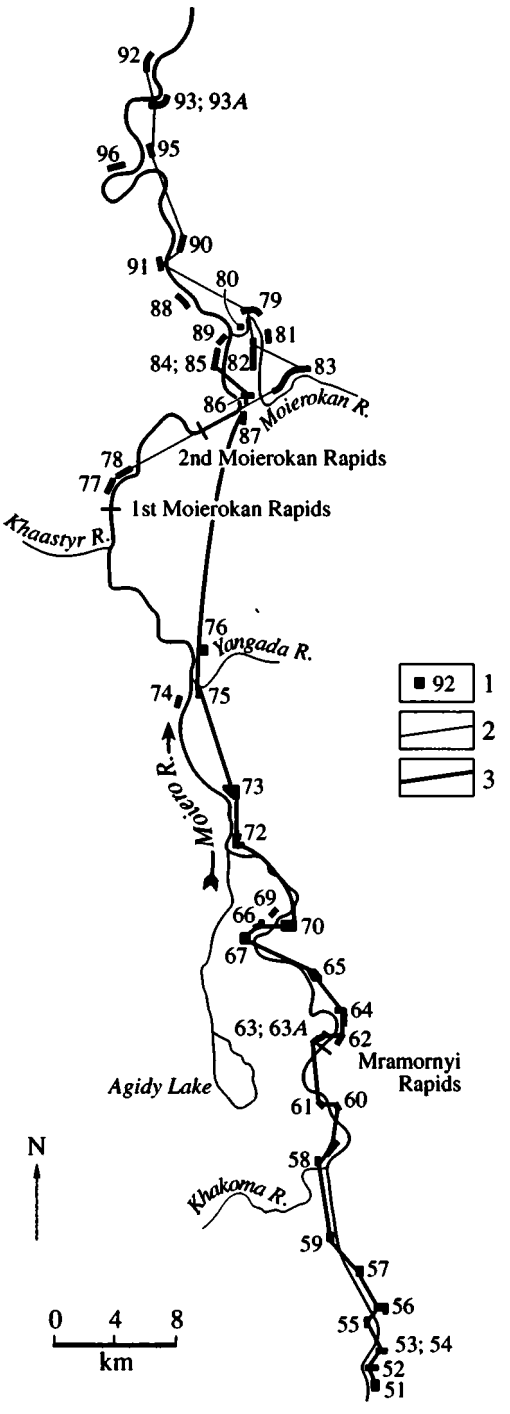

Fig. 2. Localities of Silurian sections along the Moiero River: (1) exposure numbers; (2) profile of exposures of the Moierokan and Khaastyr formations; (3) profile of exposures of the Agidy and Khakoma formations.

The formation is composed of gray terrigenous or terrigenous–carbonate rocks in its lower part and of pure carbonates in the middle and upper parts. The rock beds are 0.5 to 12.5 m thick. Black limestones and shales characteristic of the basal portion grade upward into gray platty limestones intercalated with black shales. Gray aphanic limestones of the middle interval show the knobby bedding planes. The top portion of the formation consists of gray grumous limestones. Organic remains are abundant. Dominant fossil groups are graptolites in the lower beds, trilobites in the middle interval, and brachiopods in the upper one. The formation exposures are typical of cliffs and other positive relief forms. Upper and lower boundaries of the unit are distinct. The unit is about 110 m thick and divisible into two subformations.

The **lower Moierokan Subformation** (mr_1) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 46). Its stratotype in Exposure 82 includes beds 1–16 and is 54.0 m thick. The lower beds predominantly consist of black shales. The middle part reveals the frequent alternation of shale, marl, and platty limestone beds. Limestones with knobby bedding planes are characteristic of the upper interval. Graptolites, chiolites, and small pelecypods are dominant fossils of the shale beds. In limestones, organic remains are infrequent, usually concentrated over the bedding planes. Dominant here are trilobites associated with sporadic brachiopods and gastropods. Upper limestone beds with knobby bedding planes yield mostly brachiopods and correspond to the first occurrence level of rare ostracodes, small tabulate colonies, rugose corals, bryozoans, and crinoid columnals. The subformation either rests parallel to the erosion surface of the Ordovician Dzherom Formation,

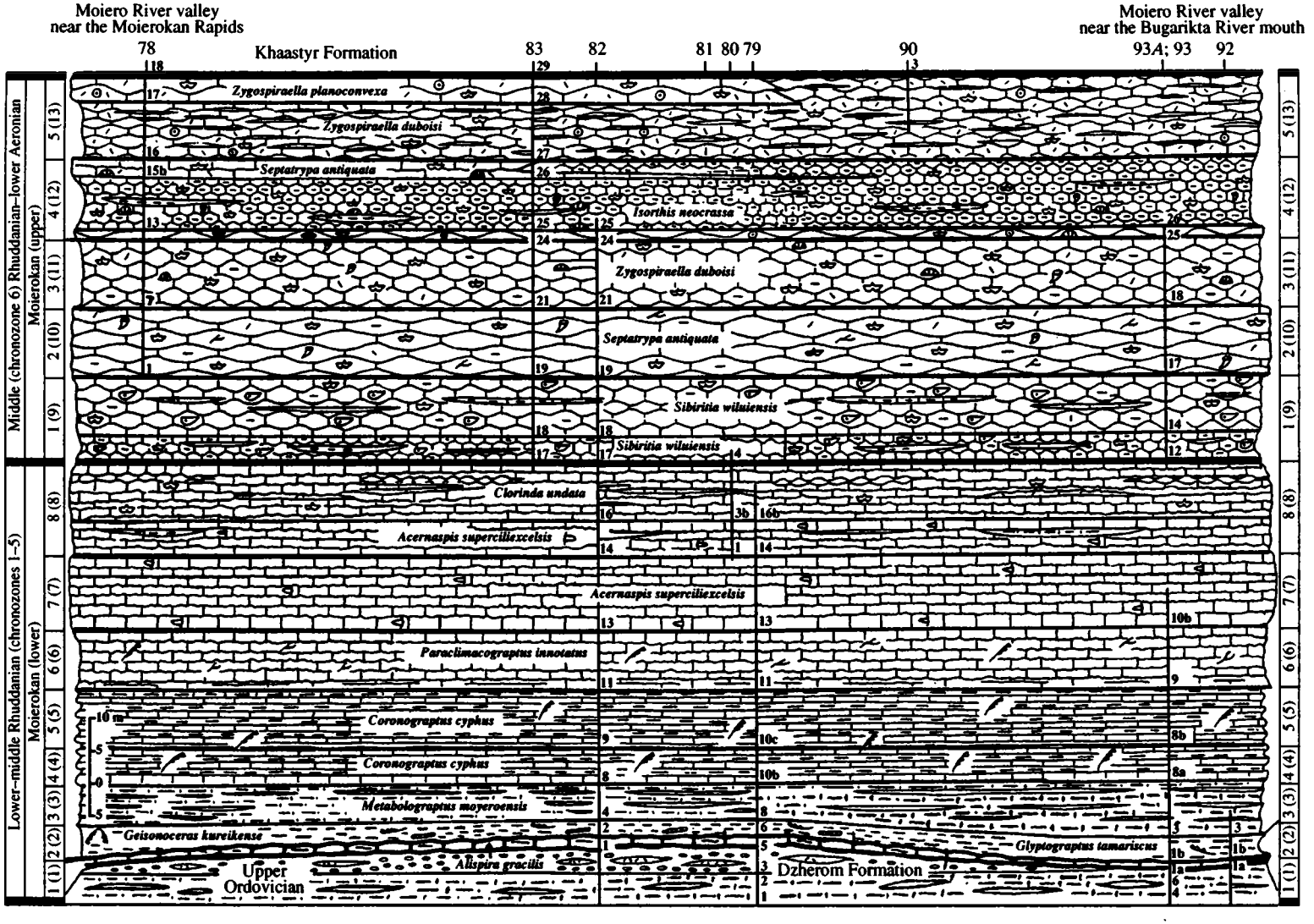

Fig. 3. Stratigraphy, lithology, and characteristic fossil groups of the Moierokan Formation (Rhuddanian–lower Aeronian), the Moiero River valley between the Bugarikta (north) and Khaastyr (south) rivers (symbols as in Fig. 8).

or overlies the latter with a minor angular unconformity. The hiatus spans an upper part of the Ashgillian Stage and locally extends down to upper Caradocian Stage. The subformation represents the stratotype of the lower Moierokan Subhorizon corresponding to the lower–middle Rhuddanian. The unit is divided into eight local chronozones.

(1) Chronozone $mr_1(1)$. Stratotype of the unit is 0.5 m thick and corresponds to Bed 82/1 composed of dark-gray fine-grained bituminous limestones. These dense platty rocks enclose pyrite nodules and show knobby bedding planes. They are intercalated with thin (1.0–1.5 cm) beaded laminae of black shale. Frequent organic remains are small in size and buried in a dispersed state; they are rare in shale intercalations. The dominant form is *Alispira gracilis.*

(2) Chronozone $mr_1(2)$. Stratotype of the unit includes beds 82/2 and 82/3 1.8 m thick together. The lower bed of black limestones yields abundant cephalopods, the dominant *Geisonoceras kureikense* included. The upper bed of dark-gray shales is rich in graptolites whose dominant form is *Glyptograptus tamariscus.*

(3) Chronozone $mr_1(3)$. The unit's stratotype 5.0 m thick corresponds to the interval of beds 82/4–82/7 composed of gray shales and light gray marl with lenticular intercalations of gray, clayey to dolomitic, fine-grained limestones displaying horizontal bedding. Rare and small, though frequent organic remains include the dominant *Metabolograptus moyeroensis.*

(4) Chronozone $mr_1(4)$. The unit's stratotype 5.0 m thick corresponds to the Bed 82/8.

(5) Chronozone $mr_1(5)$: The unit's stratotype 8.5 m thick corresponds to beds 82/9 and 82/10. The fourth and fifth chronozones represent a single geological body of gray calcareous shale enclosing frequent and thin (2–3 cm) intercalations of gray, clayey to dolomitic, fine-grained limestone, which show flat or somewhat wavy bedding planes. Rare and small organic remains are dispersed mostly over the bedding planes of shale beds. Dichotomous fucoids (1 to 18 mm in diameter) mark the bedding planes of limestone beds. The dominant form is *Coronograptus cyphus.*

(6) Chronozone $mr_1(6)$. Stratotype of the unit is 8.3 m thick and corresponds to beds 82/11 and 82/12 representing a single geological body of gray fine-grained dolomitic or slightly clayey limestones displaying small-knobby bedding planes and intercalated in the lower part with thin (1–2 cm) laminae of greenish gray clayey marl. Rare and small organic remains are dispersed. The rocks exhibit two types of fucoids: large (10 mm in diameter) decorating the bedding planes and small (up to 1.3 mm in diameter) crossing the beds. The dominant form is *Paraclimacograptus innotatus.*

(7) Chronozone $mr_1(7)$: The unit's stratotype 11.0 m thick corresponds to Bed 82/13. This geological body is of a homogeneous thin-platty structure. It consists of gray fine-grained limestone whose bedding planes are knobby. Frequent and small organic remains are concentrated at the bedding planes. The dominant form is *Acernaspis superciliexcelsis.*

(8) Chronozone $mr_1(8)$. The unit's stratotype 13.7 m thick corresponds to the interval of beds 82/14–82/16. Its lower part (beds 82/14 and 82/15) shows variable structures: thin-platty to beaded. Predominant rocks are gray to greenish gray, dense and fine-grained dolomitic limestones with a minor clay admixture. Intercalated small lentils consist of greenish gray marl, or of light-gray friable platty limestone. Rare and small organic remains from this level include the dominant *Acernaspis superciliexcelsis.* The upper Bed 82/16 is coarse-platty to thin-lenticular in structure. Gray fine-grained dolomitic limestones of the bed are intercalated with thin laminae of greenish gray marl. Small but abundant organic remains are either dispersed or clustered. The dominant form is *Clorinda undata.*

The *upper Moierokan Subformation* mr_2 was distinguished in 1979 (Tesakov *et al.*, 1979, p. 46). Its stratotype includes beds 17–23 of Exposure 82 and beds 24–28 of Exposure 83; the total thickness is 57.0 m. The unit is composed of gray limestone beds; the rock varieties can be grumous, fine-grained, dolomitic, and clayey, detrital, or bioturbated. Marly envelopes surround individual grain clusters in the rocks. The content of unsorted detritus increases upward in the section. Occasional lentils between beds consist of polydetrital limestone. Abundant, small- to medium-sized organic remains are usually dispersed. Frequent fossil groups are tabulate and rugose corals, brachiopods, ostracodes, and crinoids. The lower subformation boundary is distinct and corresponds to the first occurrence level of grumous limestones. This is the stratotype of the upper Moierokan Subhorizon corresponding in stratigraphic range to the terminal middle Rhuddanian–lower Aeronian interval. The subformation is divided into five local chronozones.

(1) Chronozone $mr_2(1)$. Its stratotype 11.8 m thick includes beds 82/17 and 82/18. The lower Bed 82/17 is "dimorphous," consisting of two rock types: gray cavernous fine-grained limestone dolomitic or slightly clayey in composition and greenish gray marl, whose lentils enclose subordinate limestone intercalations displaying the knobby bedding planes. Small- to medium-sized organic remains are abundant and occur either in the dispersed state or in clusters. The dominant form is *Sibiritia wiluiensis.* The upper, more homogeneous Bed 82/18 is composed of gray grumous limestone, which is also dolomitic and shows some clay admixture. In its upper part, there are lentils of marl. Abundant, dispersed or clustered organic remains are small. *Sibiritia wiluiensis* is again the dominant form.

(2) Chronozone $mr_2(2)$. The unit's stratotype 10.5 m thick corresponds to the interval of beds 82/19 and 82/20. Both beds consist of gray grumous fine-grained limestone showing the clay and detritus admixture. Abundant dispersed and clustered organic remains are

small to medium in size. The dominant species is *Septatrypa antiquata.*

(3) Chronozone $mr_2(3)$. Stratotype of the unit is 10.2 m thick and corresponds to beds 82/21–82/23. All beds are composed of gray grumous dolomitic limestone with laminae of greenish gray friable marl. Dispersed but abundant organic remains are small to medium in size. Their dominant form is *Calamopora alveolaris.*

(4) Cronozone $mr_2(4)$. The unit's stratotype 11.4 m thick corresponds to beds 83/24–83/26. The lower Bed 83/24 of dark-gray grumous fine-grained dolomitic to clayey limestone yields abundant organic remains small in size and predominantly buried in clusters. The dominant species is *Isorthis neocrassa.* Bed 83/25 has the diachronous upper boundary. In the north of the region (Exposure 93), the bed corresponds to the whole range of the chronozone, being overlaid in the south (Exposure 78) by the third bed of the unit. Grumous limestone of the second bed encloses laminae and small lenses of greenish gray marl. Abundant, though dispersed, organic remains of the bed are small to medium in size. *Isorthis neocrassa* is dominant amid them. The third bed (78/15b) is recognized only in the south, where it is composed of gray platty detrital limestone with wavy bedding planes. Abundant organic remains small to medium in size are buried here in a dispersed state. The dominant form is *Septatrypa antiquata.*

(5) Chronozone $mr_2(5)$. The unit's stratotype 13.0 m thick corresponds to beds 83/27 and 83/28. The lower Bed 83/27 has the diachronous upper boundary and consists of gray grumous limestone with knobby lamination. The rock encloses frequent laminae of greenish gray marl and thin lentils of dense detrital limestone showing the horizontal bedding. Small- to medium-sized fossils are abundant and buried in the dispersed or clustered state. *Zygospiraella duboisi* is dominant here. The upper Bed 83/28 of gray grumous detrital limestone is exposed only in the south of the region. It yields small abundant fossils, which are dispersed or buried in clusters. The dominant form is *Zygospiraella planoconvexa.*

The *Khaastyr Formation* (Fig. 4) was distinguished in 1979 (Tesakov *et al.*, 1979) and named after the Khaastyr River, the left tributary of the Moiero River. The composite stratotype of the unit is 129.5 m thick and corresponds to beds 18–31 of Exposure 78, 18–21 of Exposure 90, and 11–43 of Exposure 77, which have been studied in middle courses of the Moiero River and upper courses of the Kotui River. In addition, terrigenous–carbonate sediments of this formation were penetrated by boreholes Chirinda-1 and Chirinda-2. Characteristic of the formation are alternating gray members of grumous limestones and marls, both enclosing lentils of detrital material, and local authigenic breccias. Abundant fossils, usually medium in size and displaying prevalence of tabulate corals and brachiopods, are buried in the dispersed or clustered state. Coral–stromatoporoid bioherms frequently occur in basal members, and pentamerid colonies are buried within the middle interval. The well-bedded structure of the formation is responsible for the development of cristate relief forms. The lower and upper boundaries are distinct. The unit, about 130 m thick in total, is divided into two subformations.

The *lower Khaastyr Subformation* (hs_1) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 48). Its stratotype 79.5 m thick corresponds to following rock sequence: beds 18–31 (Exposure 78), 18–21 (Exposure 90), and 11–19 (Exposure 77). The unit shows an alternation of gray grumous polydetrital dolomitic limestone beds and greenish gray friable marl beds. Occurring throughout the rock sequence are frequent lentils of massive or microlaminated polydetrital limestones showing low clay content and distinct erosional contacts with the host beds of grumous limestone and marl. Another characteristic feature is the presence of coral–stromatoporoid bioherms whose size is anywhere from a few meters across to 4 × 16 m. Tabulate corals and brachiopods are dominant amid abundant fossils of variable size, which are either dispersed or buried in accumulations. The lower boundary is placed at the appearance level of bioherms and imbedded lentils of polydetrital rocks. This stratotype of the lower Khaastyr Subhorizon is about 80 m thick and corresponds to the lower part of the middle Aeronian. It is divided into four local chronozones.

(1) Chronozone $hs_1(1)$. The unit's stratotype is 20.3 m thick and corresponds to beds 78/18–78/24 composed of gray to brownish gray, grumous fine-grained and clayey limestone displaying knobby bedding and intercalated with bioherms and lentils of gray polydetrital limestone. Taeniate bryozoans are dominant fossils of the limestone lentils. Bioherms are predominantly composed of corals. Tabulates and brachiopods are dispersed in grumous limestone beds. The dominant species is *Quadralites quadratus.*

(2) Chronozone $hs_1(2)$. The unit's stratotype is 20.3 m thick and corresponds to beds 78/25–78/31 and 90/18–90/21. Its lower beds 78/25–78/27 are composed of gray grumous fine-grained detrital limestone with lentils of bryozoan–crinoid limestone and greenish gray marl. Abundant fossils small to medium in size are dispersed or occur in accumulations. *Stegerhynchus decemplicatus duplex* is dominant in their burials. The next Bed 78/28 of greenish gray friable marl encloses thin frequent lentils of organogenic detrital limestone. Small- to medium-sized fossils are found in the dispersed state and in local accumulations at the bedding planes. Their dominant form is *Stegerhynchus decemplicatus duplex.* In the overlying beds 78/29–78/31 of gray grumous detrital and clayey limestones, both intercalated with thin lentils of polydetrital limestone and marl, organic remains of small and medium size are again abundant. *Calamopora alveolaris* is dominant amid dispersed fossils and in their accumulations. The

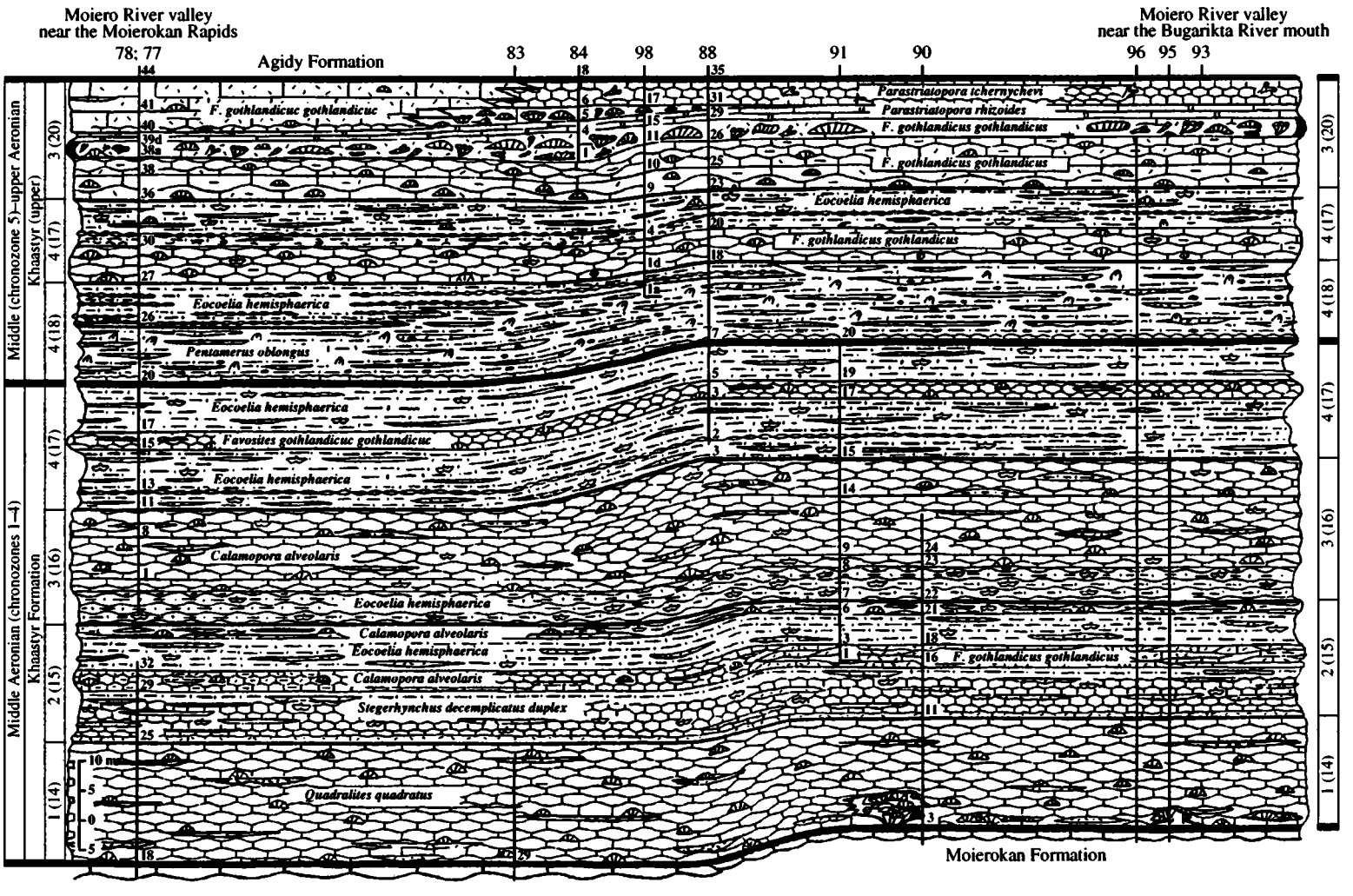

Fig. 4. Stratigraphy, lithology, and characteristic fossil groups of the Khaastyr Formation (Middle–upper Aeronian), the Moiero River valley between the Bugarikta (north) and Khaastyr (south) rivers (symbols as in Fig. 8).

upper beds 90/18–90/21 consist of greenish gray to green marls and lentils of gray massive polydetrital limestone enclosing autoclastic breccias. Abundant organic remains variable in size are concentrated over the bedding planes. The dominant form is *Eocoelia hemispherica.*

(3) Chronozone $hs_1(3)$. The stratotype of the unit (18.0 m thick) corresponds to beds 90/22, 90/23, and 77/1–77/10. The lower bed 90/22 of greenish gray grumous marl encloses thin lentils of gray massive detrital limestone. Small to large fossils from marl are usually dispersed; in limestone lentils, they are small and occur in accumulations. *Eocoelia hemispherica* is a dominant form of the bed. The overlying gray beds 90/23 and 77/1–77/10 of grumous fine-grained clayey to detrital limestone are intercalated with rare thin lentils of massive detrital limestone and greenish gray marl. The beds yield abundant, small- to medium-sized fossils dispersed or clustered in the rocks. *Calamopora alveolaris* is dominant amid other taxa.

(4) Cronozone $hs_1(4)$. The unit's stratotype is 21.0 m thick and spans the interval of beds 77/11–77/19. The lower beds 77/11–77/14 of marl and green splintery shale enclose rare thin lentils of gray massive polydetrital limestone with uneven bedding planes. Occasional lentils consist of gray grumous limestone. Small- to medium-sized fossils are abundant only in limestone lentils. *Eocoelia hemispherica* is the dominant species. The overlying gray beds 77/15 and 77/16 of grumous fine-grained dolomitic limestone with abundant detritus show accumulations of organic remains. *Favosites gothlandicus gothlandicus* is dominant amid the latter. The upper beds 77/17–77/19 of green splintery shales are intercalated with rare thin lentils of massive detrital limestone with rough bedding planes. Rare and small fossils, the dominant *Eocoelia hemispherica* included, occur mainly at these planes.

The **upper Khaastyr Subformation** (hs_2) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 48), and beds 20–43 in the Exposure 77 represent its stratotype 50.0 m thick. The subformation is widespread everywhere in the Moiero River vicinity. It shows alternation of green, greenish gray, and gray marl beds intercalated with gray lentils and interbeds of detrital or biogenic grumous limestones. Excess dolomite content, abundant corals, and fossil pentamerid colonies are characteristic of this bed sequence. Abundant fossils are medium to large in size; they are dispersed in the rocks or buried as accumulations. The apparent lower boundary corresponds to the level of closely spaced interbeds of detrital limestones, which is also marked by occurrence of authigenic breccias. This is the type unit of the upper Khaastyr Subhorizon spanning the terminal middle–upper Aeronian interval. The subhorizon is about 50.0 m thick and includes three local chronozones.

(1) Chronozone $hs_2(1)$. Stratotype of the chronozone corresponds to beds 77/20–77/26, is 16.5 m thick, and includes two diachronic subunits. In the north of the region, the lower subunit spans the entire range of the chronozone. In the south, the upper one above the diachronic boundary overlies it. The lower subunit (beds 77/20–77/25) is composed of marls and greenish gray splintery shales intercalated with frequent thin lentils of gray polydetrital, locally clayey and grumous limestones. Abundant small- to medium-sized fossils occur in limestones only, where they are dispersed or form accumulations. *Pentamerus oblongus* is dominant in this case. The upper Bed 77/26 of marl and greenish gray shale encloses frequent and rather thick lentils of gray grumous polydetrital limestone and rare thin lenses of detrital limestone. Fossils of the same size range are, as a rule, in limestones. *Eocoelia hemispherica* is dominant amid them.

(2) Chronozone $hs_2(2)$. Beds 77/27–77/35 13.6 m thick in total represent stratotype of the chronozone. The lower bed sequence (77/27–77/29) consists of gray grumous fine-grained detrital limestone dolomitic in composition and enclosing tiny, greenish gray lenses of nodular limestone. Rare and dispersed fossils are small to medium in size. The dominant species is *Favosites gothlandicus gothlandicus.* The upper beds 77/30–77/35 are composed of marls and greenish gray splintery shales with thin and frequent lenticular intercalations of gray detrital limestone. Accumulations of small- to medium-sized fossils are buried on bedding planes of limestone lentils. *Eocoelia hemispherica* is dominant amid other taxa.

(3) Chronozone $hs_2(3)$. The unit's stratotype 20.0 m thick corresponds to beds 77/36–77/43 divided into five subunits. The lower one (beds 77/36–77/38) is composed of gray to greenish gray grumous fine-grained polydetrital clayey-dolomitic limestones. Abundant large and smaller organic remains are either dispersed, or occur in stratal accumulations. *Favosites gothlandicus gothlandicus* is the dominant taxon of the subunit. Overlying beds 77/39a–77/39c of black platty clayey-detrital limestone yield abundant corals and the dominant *Favosites gothlandicus gothlandicus.* The third subunit (beds 77/39c–77/39e) consists of greenish gray fine-laminated dolomitic limestone enclosing lentils of grumous clayey lime. It yields the dispersed or clustered medium-sized fossils, the dominant *Parastriatopora rhizoides* included. The fourth subunit includes gray beds (77/40–77/43) of medium- and thick-platty, locally grumous limestones with thin massive detrital lentils. *Favosites gothlandicus gothlandicus* is the dominant taxon of dispersed or clustered organic remains buried in these rocks. The uppermost beds 84/5–84/7 of gray grumous limestone are rich in coral colonies and yield *Parastriatopora tchernychevi* as the dominant taxon.

The *Agidy Formation* (Fig. 5) is distinguished in 1979 (Tesakov *et al.*, 1979, p. 49) and named after its eponymous lake. Its stratotype, 84.5 m thick, currently corresponds to beds 8–56 distinguished in exposures 84–87. Originally (Tesakov *et al.*, 1979, p. 49, Fig. 2),

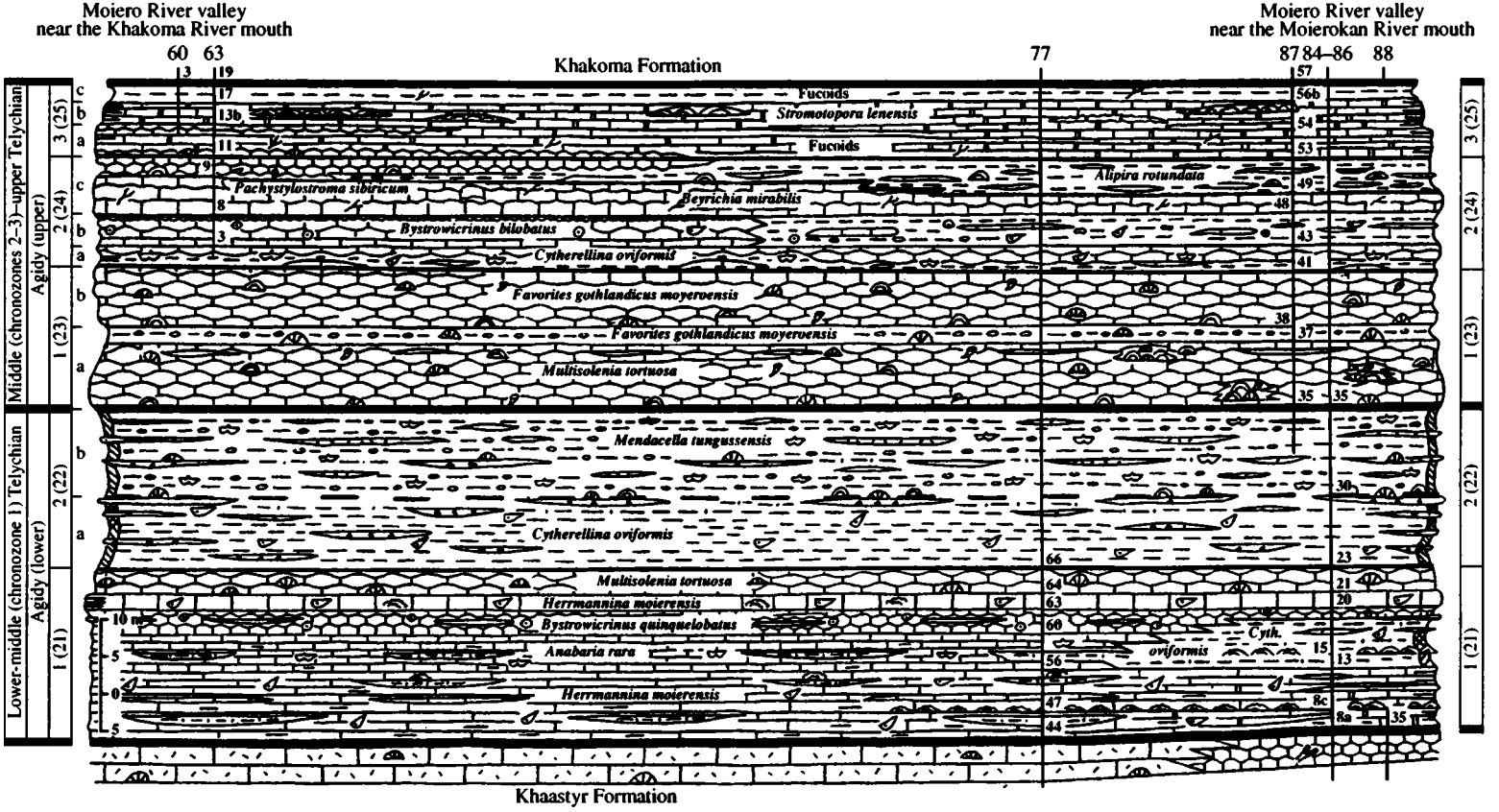

Fig. 5. Stratigraphy, lithology, and characteristic fossil groups of the Agidy Formation (Telychian), the Moiero River valley between the Moierokan (north) and Khakoma (south) rivers (symbols as in Fig. 8).

we considered as reference sections those of partially isolated exposures 84–86, 77, and 63. Afterward (Tesakov *et al.*, 1985, p. 85, Fig. 22), only exposures 84–86 were deemed to fit the stratotype definition, but now the bed-by-bed correlation between exposures 63 and 87 is established. Accordingly, the entire Exposure 87 is attributed to the upper Agidy Subformation (but not to the Khakoma Formation as it was done in 1985), and the almost uninterrupted section of exposures 84–87 can be taken at present for the Agidy Formation stratotype. The rocks in question are widespread in middle courses of the Moiero River and in the source area of the Kotui River. They are recovered as well by boreholes Chirinda-1 and Chirinda-2. The formation includes three lithological units: the lower one of platty and grumous limestones, middle beds of marl, and the greater upper part consisting of grumous limestone. In general, the formation is rich in fossils, but their abundance rate is highly variable and dependent on the rock lithology. Fossils are small to medium in size, buried either in the dispersed state, or in accumulations and biohermal assemblages. The rocks are well bedded, and those of the upper part form the positive relief forms. The formation thickness is about 84 m. The bed sequence is subdivided into two subformations.

The **lower Agidy Subformation** (ag_1), in its new range, is characterized here for the first time. Rocks of the unit are widespread everywhere in the Moiero River region. They represent two lithologic members. The lower one includes beds of gray, platty to foliated dolomitic limestones and variegated marls, which are intercalated with rare interbeds of gray grumous limestone. Rather frequent, but small organic remains (predominantly of ostracodes and brachiopods) are irregularly distributed across the bed sequence. The upper member is composed of green marl beds and gray lentils of detrital limestones. Accumulations of small- to medium-sized organic remains occur in limestone beds only. Ostracodes and brachiopods are characteristic fossils of marl beds, and corals are typical of limestones. Authigenic breccias are distinctive components of the member. The subformation is the type unit of the lower Agidy Subhorizon corresponding in range to the lower–basal middle Telychian. The subhorizon is about 42 m thick and includes two local chronozones.

(1) Chronozone $ag_1(1)$. Its stratotype 21.6 m thick corresponds to beds 84/8–84/15 and 85/16–85/22. The chronozone is divided into five subunits, the lower of which (beds 84/8–84/12) has the upper boundary distinct in the north of the region (Exposure 84) and uncertain in the southern sequence of platty limestone (Exposure 77). Predominant rocks of the beds are foliated to thin-platty dolomitic limestones that are gray, greenish, yellow-gray, and black in color; these rocks enclose rare interbeds of gray massive limestone and sporadic lentils of green marl. Fossils are rare in general, though some beds yield abundant ostracodes. *Hermannina moierensis* is the dominant taxon. Boundaries of the next subunit (beds 84/13–84/15 and 85/16–85/18) are distinctly diachronous. Principal rocks here are green, red, greenish gray, and variegated marls intercalated with subordinate lentils of gray detrital, irregularly bedded limestone. Stromatolitic biostromes are characteristic components of the beds. Other fossils, the dominant *Cytherellina oviformis* included, are small in size and buried individually or in accumulations. In the south of the region, these beds grade into beds 77/56–77/59 of the third subunit consisting of gray platty limestones intercalated with rare lentils of greenish gray marl. *Anabaria rara* is dominant at this level. The fourth subunit (Bed 85/19) is exposed everywhere in the region. In the north, it is slightly thinner than in the south. Gray to black grumous limestone of the bed encloses stratified limestone lentils and shows accumulations of crinoids, brachiopods, and ostracodes, among which *Bystrowicrinus bilobatus* is dominant. The overlying Bed 86/20 of gray thick-platty limestone yields abundant stromatolite calypters and other small dispersed or accumulated fossils, the dominant *Herrmannina moierensis* included. The uppermost sixth subunit of gray grumous limestone beds 86/21 and 86/22 bears small- to medium-sized dispersed fossils, whose dominant taxon is *Multisolenia tortuosa.*

(2) Chronozone $ag_1(2)$. The unit's stratotype 20.9 m thick corresponds to beds 86/32–86/34. Main rock types in this case are greenish friable shales and marls enclosing clots of gray limestone. Rare, small and dispersed organic remains in these rocks are mainly represented by ostracodes and brachiopods. Subordinate, usually small lentils of variable lithologic composition often consist of gray detrital limestone and autoclastic breccias. They enclose mass accumulations of brachiopods, ostracodes, and crinoids. Rare larger lentils of gray grumous limestone are rich in corals and bryozoans. *Cytherellina oviformis* is dominant in the lower subunit *a*, and *Mendacella tungussensis* is a characteristic taxon of the upper subunit *b*.

The *upper Agidy Subformation* (ag_2) has stratotype corresponding, according to recent data, to beds 87/35–87/56, whose rocks are exposed throughout the Moiero River region. The subformation consists of alternating beds of platty grumous and foliated limestones intercalated with rare interbeds of greenish gray marl. The basal beds enclose bioherms and abundant medium- to large-sized fossils occurring individually or in accumulations. Middle limestone beds yield dispersed small-shelled fauna. The upper interval bears only brachiopods and ostracodes. The stratotype is taken for typical of the upper Agidy Subhorizon corresponding to the terminal middle–upper Telychian interval. The subformation is 42 m thick and divided into three local chronozones.

(1) Chronozone $ag_2(1)$. Beds 87/35–87/40 as thick as 17.6 m in total represent stratotype of the chronozone. Lower beds 87/35 and 87/36 mainly consist of gray grumous fine-grained detrital limestone enclosing

coral-stromatoporoid bioherms and lentils of gray massive limestone. Abundant organic remains of medium and large size are dispersed in detrital limestones or form accumulations in limestone lentils. *Multisolenia tortuosa* is dominant. The middle Bed 87/37 of greenish gray marl with clots of clayey limestone yields dispersed fossils that are small and medium in size, the dominant *Favosites gothlandicus moyeroensis* included. Upper beds 87/38–87/40 are composed of gray grumous fine-grained clayey-detrital limestone. Abundant, medium- to large-sized fossils are usually dispersed. Their local accumulations transform the rocks into biomorphic varieties. *Favosites gothlandicus moyeroensis* is again the dominant taxon.

(2) Chronozone $ag_2(2)$. Its stratotype14.8 m thick corresponds to beds 87/41–87/52 divided into five lithologic units. The lower beds 87/41–87/47 of greenish gray friable marl enclosing limy clots are intercalated with two types of lentils: larger ones composed of gray grumous fine-grained clayey limestone rich in organic remains and smaller lentils of brownish gray platty limestone, where fossils are less abundant. *Cytherellina oviformis* represents the dominant taxon of the beds. Beds 63/3–63/7 next in the sequence consist of grumous limestone bearing *Bystrowicrinus bilobatus*. The overlying Bed 87/48 is thinner in the north than in the south. Gray dolomitic limestone of the bed shows wavy bedding planes and interlaminations of detrital limestone and marl. *Beyrichia mirabilis* is dominant at this level. The fourth lithologic unit (beds 87/49–87/52) characteristic of northern exposures is pinching out toward the south. It consists of greenish gray marl enclosing nodules and lenses of gray detrital limestone, which encloses autoclasts and shows wavy or knobby bedding. *Alispira rotundata* is the dominant taxon here. The fifth (southern) lithologic unit (beds 63/9–63/11) of gray grumous clayey limestone is rich in small- to medium-sized fossils buried individually or in accumulation. The dominant taxon is *Pachystylostroma sibiricum*.

(3) Chronozone $ag_2(3)$. Stratotype of the chronozone is 9.4 m thick and corresponds to beds 87/53–87/56. The lower Bed 87/53 consists of gray thin-platty dolomitic limestone. In southern exposures of the region, limy clots from the upper portion of the bed yield rare but diverse fossils, whereas the matrix of the rock shows only mud eaters' tracks. The middle beds 87/54–87/56a of gray laminated to foliated dolomitic limestones show lenticular intercalations of dark gray platty limestone bearing microphytoliths and enclose thin stromatolite biostromes. Organic remains, the dominant *Stromatopora lenensis* included, occur at particular stratal levels only. The upper Bed 87/56b consists of greenish gray shale interlaminated with tiny lenses of limestone and dolomite. Only impressions of ostracode shells are recorded here.

The **Khakoma Formation** (Fig. 6) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 50) and named after the Khakoma River, the left tributary of the Moiero River. The respective stratotype includes beds 60/3, 60/4, 58A/4–58A/8, 58/2–58/22, 64/12–64/18, and 66/1b–66/31 studied prior to 1979 in the reference exposures 60, 58A, 58, and 64. Later (Tesakov *et al.*, 1985, p. 97), we distinguished parastratotypes of the lower (Exposure 87) and upper (Exposure 66) subformations, which have been considered afterward (Tesakov *et al.*, 1992, Fig. 2) as stratotypes of these subdivisions. However, our recent data show that the Exposure 87 is a stratigraphic analogue of the Exposure 63 exemplifying beds of the Agidy Formation. This complex investigation history is a consequence of the highly variable facies composition of the Khakoma Formation, whose rocks are exposed in the steep 40-m-high cliffs making it inconvenient to conduct detailed biostratigraphic investigation and correlation analysis. Rocks of the formation are widespread in middle courses of the Moiero River and in headwaters of the Kotui River; boreholes Chirinda-1 and Chirinda-2 have also recovered them. Characteristic of the formation lithology are coral-algae-stromatoporoid bioherms and biostromes, some up to 8 m high. Space between these biogenic structures is filled with gray limestones of variable structure and composition. These fine- to thick-platty and foliated rocks are fine- to coarse-grained, rich in or free of detrital material, and biogenic or abiogenic in origin. Similar features are characteristic of dolomites widespread in the Khakoma bed sequence. In addition, the sequence includes lentils of clastic limestones, authigenic breccias, marl lenses, and cherty nodules. Abundant organic remains are medium to large in size and occur individually or in accumulations. In bioherms and biostromes, they are the framebuilding fossils. Most characteristic fossil groups are stromatoporoids, tabulates, rugose corals, ostracodes, gastropods, nautiloids, and crinoids. In general, the formation is of a mosaic structure. Its rocks are exposed in the positive relief forms. The lower and upper boundaries are distinct. The whole unit, about 82 m thick, is divided in two subformations.

The **lower Khakoma Subformation** (hk_1) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 51). Its stratotype 42.0 m thick corresponds to beds 60/3, 60/4, 58A/2–58A/8, 58/2–58/22, and 64/12–64/18. General features are identical to those of the formation. The subformation represents the stratotype of the lower Khakoma Subhorizon spanning the time range of the Sheinwoodian Stage. It includes five local chronozones.

(1) Chronozone $hk_1(1)$. Stratotype of the chronozone is 9.2 m thick, cyclic in structure, and corresponds to beds 60/3, 60/4, 58A/2, and 58A/3. The lower beds 60/3 and 60/4 are of variable lithology. Gray platty limestones, which show wavy lamination and yield abundant Coelenterata, represent the main rock type. They enclose bioherms up to 5–7 m high and stratal or elongated-lenticular biostromes. *Labechia condensa* is the dominant taxon of these beds. The overlying middle

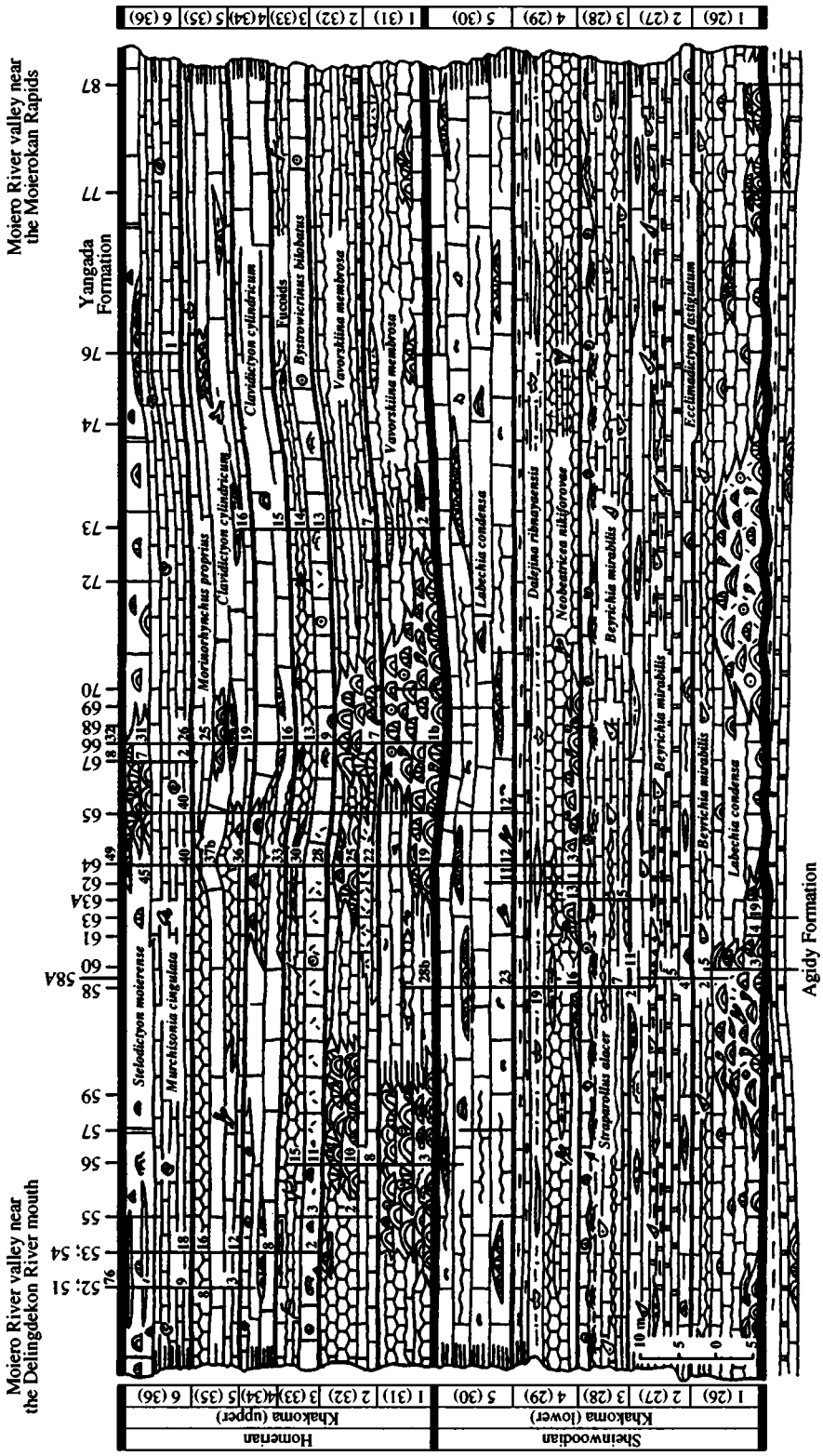

Fig. 6. Stratigraphy, lithology, and characteristic fossil groups of the Khakoma Formation (Wenlockian), the Moiero River valley between the Moierokan (north) and Delingdekon (south) rivers (symbols as in Fig. 8).

Bed 58A/2 of gray grumous fine-grained limestone bears small rare fossils dispersed in the rock. Dominant here is *Beyrichia mirabelis*. The upper Bed 58A/3 of light gray, thin-platty to foliated limestones shows ripple marks on bedding planes. It yields rare and small fossils, the dominant *Herrmannina nana* included.

(2) Chronozone $hk_1(2)$. Stratotype of the chronozone is 8.3 m thick and corresponds to beds 58A/4–58A/8 and 58/2–58/6. Its lithocoenosis consists of the following rock types: light to dark gray, medium- or thin-platty dolomitic-clayey limestones, the stromatoporoid, stromatolitic, oncolitic, and detrital rock varieties included; greenish gray marls; green friable shales; limestone lentils with authigenic breccias; and small-pebbled conglomerates. Rare and small organic remains are usually dispersed in the rocks and occur locally in accumulations. *Beyrichia mirabilis* is the dominant taxon.

(3) Chronozone $hk_1(3)$. Stratotype of the chronozone is 6.2 m thick and corresponds to beds 58/7–58/15. The lower beds 58/7–58/14 are composed mainly of gray thin-platty limestones intercalated with large lentils of gray grumous clotted-detrital limestone and small lentils of greenish gray friable dolomitic marls. Dispersed or clustered fossils, amid which *Beyrichia mirabilis* is dominant, are rather abundant. The overlying Bed 58/15 of banded, gray to black grumous and well-bedded limestones encloses thin stromatolite build-ups. Stratal accumulations of abundant organic remains are typical of the bed. The dominant taxon is *Straparollus alacer.*

(4) Cronozone $hk_1(4)$. Stratotype of the chronozone is 8.5 m thick and corresponds to beds 58/16–58/22. The lower beds 58/16–58/19 consist of several rock types. Gray to greenish gray, platty clayey-detrital limestones are intercalated here with lentils of green marl and extremely clayey limestone. The rocks also enclose biostromes. Irregularly dispersed organic remains are abundant and small to medium in size. *Neobeatricea nikiforovae* is the dominant taxon. The middle Bed 58/20 of banded gray marl encloses thin lentils of gray limestone and dolomite. Rare and small organic remains show predominance of *Dalejina ribnayaensis.* The upper beds 58/21 and 58/22 are composed of gray thick-platty to foliated dolomites. Fossils are small-sized and occasional here. The dominant taxon is the same.

(5) Chronozone $hk_1(5)$. Stratotype of the chronozone is 9.8 m thick and corresponds to beds 64/12–64/18. Lithologic rock varieties are gray thick- to medium-platty limestones barren of fauna, organogenic detrital limestones, and lentils of gray grumous limestones. These rocks envelop occasional bioherms and cherty segregations. Large and medium-sized fossils are abundant only in grumous limestones, where they are buried individually or in accumulations. Their dominant form is *Labechia condensa.*

The **upper Khakoma Subformation** (hk_2) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 52). A stratotype of this subdivision corresponds to beds 66/1b–66/31, whose sequence is 40.0 m thick. General lithological features are similar to those of the formation. The occurrence level of thick-platty limestone beds with numerous thick bioherms marks the boundary between the lower and upper subformations. The subdivision represents the stratotype of the upper Khakoma Subhorizon corresponding in time range to the Homerian Stage. Six local chronozones are distinguished in the subhorizon.

(1) Chronozone $hk_2(1)$. Stratotype of the chronozone is 8.0 m thick and corresponds to beds 66/1b–66/6. Lithologic rock varieties are gray to light gray and yellowish banded thick- to medium-platty detrital limestones displaying fine- to coarse-grained texture and intercalated with small lentils of gray grumous and massive detrital limestones. These rocks enclose large coral-stromatoporoid bioherms. Large and medium-sized fossils are distributed irregularly and form framework of bioherms. The dominant taxon is *Yavorskiina membrosa.*

(2) Chronozone $hk_2(2)$. Stratotype of the chronozone is 6.1 m thick and corresponds to beds 66/7–66/8. The rock varieties are gray medium-platty grumous and detrital limestones with rather large coral-stromatoporoid bioherms and biostromes. Fossils are rare in limestones but abundant in bioherms and biostromes, where they are large. *Yavorskiina membrosa* is again the dominant taxon.

(3) Chronozone $hk_2(3)$. Stratotype of the chronozone is 5.4 m thick and corresponds to beds 66/9–66/14. Rocks of the lower beds 66/9–66/12 are gray to brownish gray platty crinoidal and stromatolitic detrital limestones somewhat clayey in composition. Rare fossils are distributed at random across the beds. Their dominant form is *Bystrowicrinus bilobatus.* The upper beds 66/13 and 66/14 of gray grumous limestones show abundant fucoids.

(4) Chronozone $hk_2(4)$. Stratotype of the chronozone is 4.8 m thick and corresponds to beds 66/15–66/18. Rock varieties of the beds are gray to light gray, thin-platty to foliated limestones intercalated with lentils of orange biomorphic dolomites and gray grumous limestones. Fossils are abundant only in biomorphic lentils. The dominant taxon is *Clavidictyon cylindricum.*

(5) Chronozone $hk_2(5)$. Stratotype of the chronozone is 8.2 m thick and corresponds to beds 66/19–66/25. Rock varieties of the unit are light- to dark-gray limestone, mostly foliated with single small-sized organic remains and subordinate lentils of light-gray limestone devoid of fossils or dark-gray biogenic limestone. In the southern part of the considered area, there are large lentils of gray grumous limestone enriched in medium-sized organic remains. The dominant form is *Clavidictyon cylindricum.*

(6) Chronozone $hk_2(6)$. Stratotype of the chronozone is 8.0 m thick and corresponds to beds 66/26–66/31. Lithologic rock varieties in the lower beds 66/26–66/30 are black thick-platty limestones, with rare lentils of dark gray grumous and brownish gray thin-platty limestones. Dispersed fossils are locally concentrated in stratal accumulations. The dominant taxon is *Murchisonia cingulata.* The uppermost Bed 66/31 consists of black thick-platty limestone intercalated with lentils of stromatolitic or detrital limestones and with coral-stromatoporoid bioherms. *Stelodictyon moierense* is dominant at this level.

The Kureika River (Turukhansk) type area of the Tukal Horizon corresponding to the Gorstian Stage is situated near the western margin of the Siberian platform (Fig. 1). Profile of exposures of the Tukal Formation, which have been studied along the river, is shown in Fig. 7.

The **Tukal Formation** (Fig. 8) was distinguished in 1979 (Tesakov *et al.*, 1979, p. 83) and named after the Tukal tribe of Tungusic people. Stratotype of the formation is 67.5 m thick and corresponds to beds 11/3–11/17, 3/2–3/30, 2/1–2/18, 1/1–1/16, 4/2–4/14, and 10/28–10/40. Three exposures on the left and right-hand side of the Kureika River reveal all rock varieties of the unit (Tesakov *et al.*, 1980, Figs. 14 and 15), whose beds are cut by numerous faults, and each faulted block is numbered individually. Rocks of the formations are exposed in the Kureika and Nizhnyaya Tunguska river basins, where many boreholes have also recovered them. The formation is diverse in lithological composition, and its rocks display highly variable textural and structural features. Remarkable components of the formation are detrital, oolitic, algal, stromatolitic, and nodular limestones, and also marls with lime nodules and dolomitic marls. Bioturbation, fucoids, exsiccation cracks, and authigenic brecciation are characteristic of the rocks. Organic remains are represented mainly by ostracode–gastropod, brachiopod, and algal assemblages. The unit divided into two subformations is about 65 m thick.

The **lower Tukal Subformation** (tk_1) was distinguished in its current range in 1992 ((Tesakov *et al.*, 1992, p. 82, Fig. 9). The stratotype of the subformation is 27.0 m thick and corresponds to beds 11/3–11/17 and 3/2–3/28. Initially (Tesakov *et al.*, 1980, p. 79, Fig. 24), we assumed a much greater thickness of this subdivision, but in practice, the idea turned out to be misleading. Principal rock types are dark gray platty limestones, frequently brecciated, bearing stromatolites, and intercalated with beds of gray marl enclosing lime nodules, and also with lentils of authigenic breccias. Small- to medium-sized organic remains are abundant only in particular beds. Predominant fossil groups are ostracodes, brachiopods, gastropods, and crinoids. The subformation represents the type unit of the lower Tukal Subhorizon corresponding in time range to the

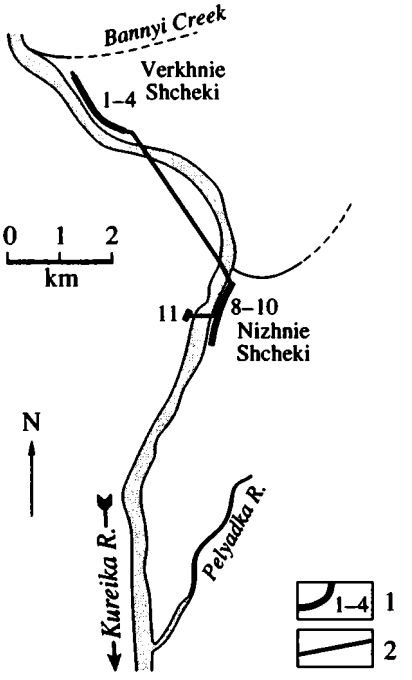

Fig. 7. Localities of Silurian sections along the Kureika River, Turukhansk area: (1) exposure numbers; (2) profile of exposures of the Tukal Formation.

lower Gorstian Substage. Four local chronozones are distinguished within the subdivision.

(1) Chronozone $tk_1(1)$. Stratotype of the chronozone is 7.9 m thick and corresponds to beds 11/3–11/13. The lower beds 11/3–11/10 of gray to dark-gray platty organogenic-detrital limestones yield microphytoliths and are intercalated with lentils of gray stromatolitic limestone and brownish gray marl. Small fossils are dispersed or concentrated only in particular lentils. *Eukloedenella kureikensis* and stromatolites represent dominant taxa. The overlying beds 11/11–11/13 are banded, consisting of dark-gray dolomitic marl, and intercalated with tiny lenses of pure gray, clayey black, and detrital dark-gray limestones. Organic remains, the dominant *Eukloedenella kureikensis* included, are concentrated only in intercalations of detrital limestone.

(2) Chronozone $tk_1(2)$. Stratotype of the chronozone is 6.4 m thick and corresponds to beds 11/14–11/17. The beds are composed of dark gray, thick- to thin-platty oolitic and stromatolitic limestones. Stromatolites and *Bystrowicrinus bilobatus* are dominant amid rare fossils of the beds.

(3) Chronozone $tk_1(3)$. Stratotype of the chronozone is 6.2 m thick and corresponds to beds 3/2–3/13. The lower beds 3/2–3/12 of gray marl with lime clots enclose lentils of gray thin-platty stromatolitic limestone. Small and abundant fossils are buried individually or in accumulations. *Hyattidina acutisummitatus* is dominant among other taxa. The upper Bed 3/13 of dark-gray massive polydetrital-clayey limestone yields

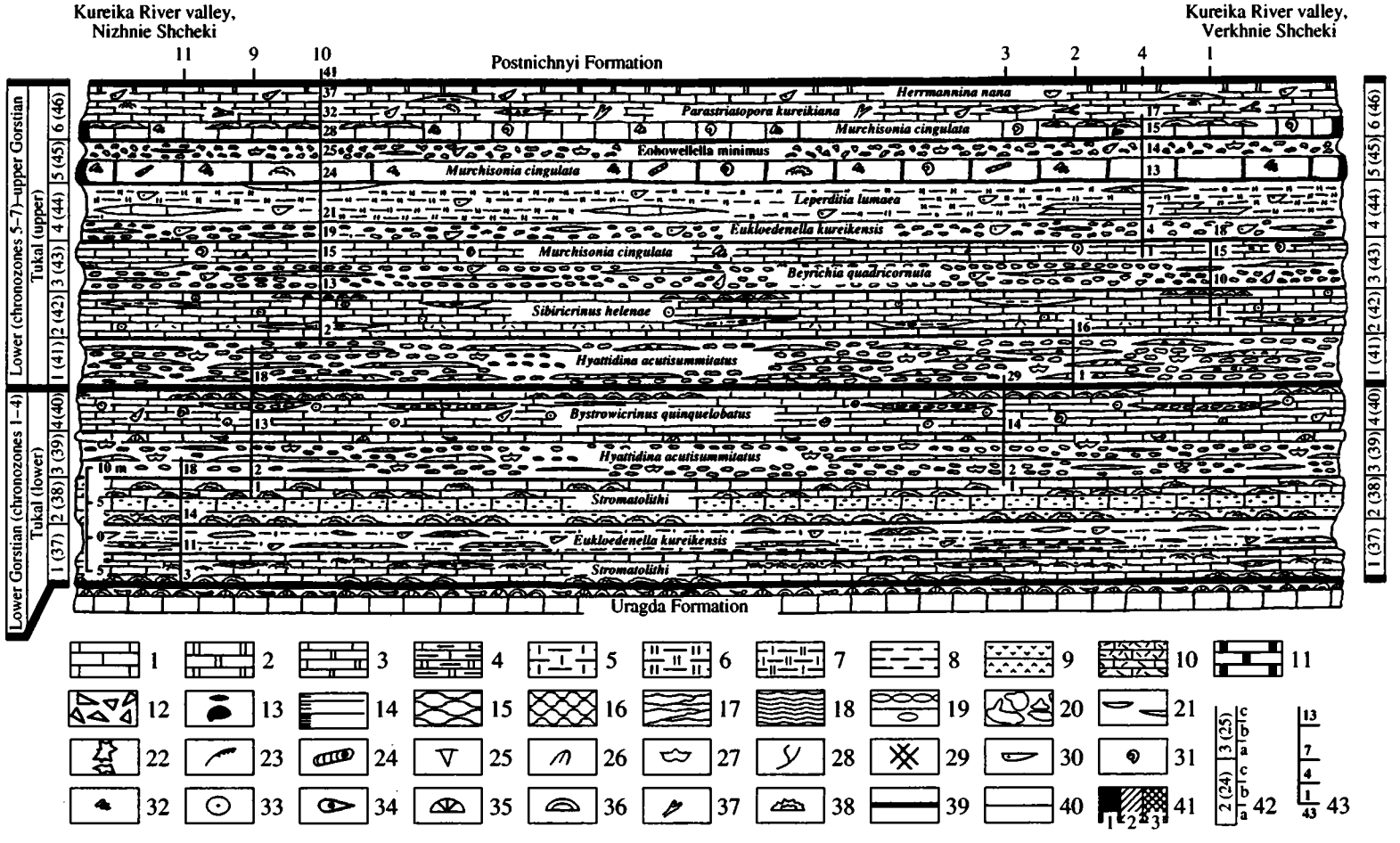

Fig. 8. Stratigraphy, lithology, and characteristic fossil groups of the Tukal Formation (Gorstian), the Kureika River valley between the Verkhnie (north) and Nizhnie (south) Shcheki localities: (1) limestone; (2) dolomite; (3) dolomitic limestone; (4) clayey limestone or dolomite; (5) marl; (6) domerite; (7) calcareous-dolomitic marl; (8) shale; (9) gypsum or anhydrite; (10) detrital limestone; (11) bituminous limestone; (12) authigenic breccia; (13) cherty nodules; (14) fine lamination; (15) coarse grumous structure; (16) fine grumous structure; (17) bedded structure; (18) knobby bedding planes; (19) nodular interbeds; (20) lime nodules; (21) lentils; (22) caverns; (23) graptolites; (24) cephalopods; (25) trilobites; (26) pentamerids; (27) brachiopods; (28) mud eaters' tracks; (29) bryozoans; (30) ostracodes; (31) flattened gastropods; (32) three-dimensional gastropods; (33) crinoids; (34) rugose corals; (35) tabulates; (36) stromatoporoids; (37) branching tabulates and stromatoporoids; (38) stromatolites; (39) boundaries between formations and subformations; (40) boundaries of chronozones; (41) dark (1), greenish (2), and variegated (3) rock colors; (42) scale of chronozones, numbers of standard and regional (in brackets) chronozones are indicated and symbols a, b, and c mark subdivisions of local chronozones; (43) exposure and borehole sections with numbered beds, whose base marks boundary between lithologic subdivisions.

Subalveolites subulosus dominant among other small or medium-sized fossils dispersed in the bed.

(4) Chronozone $tk_1(4)$. Stratotype of the chronozone is 6.4 m thick and corresponds to beds 3/14–3/28. Gray to dark-gray thin- and medium-platty limestones of the beds are intercalated with large lentils, which consist of gray stromatolitic limestone and dark-gray marl enclosing limy clots and minor lenses of detrital limestone, and also with thin lentils composed of gray to dark gray marl and dark gray limestone bearing autoclasts. Small fossils are dispersed or accumulated only in limestones. *Bystrowicrinus quinquelobatus* is the dominant taxon.

The **upper Tukal Subformation** (tk_2) is described here for the first time in the new revised stratigraphic range. Stratotype of the subformation is 40.5 m thick and corresponds to beds 3/29, 3/30, 2/1–2/18, 1/9–1/16, 4/2–4/14, and 10/28–10/40. Earlier, the subformation included an additional upper member (Tesakov *et al.*, 1979, p. 85; Tesakov *et al.*, 1980, p. 79, Fig. 24; Tesakov *et al.*, 1992, p. 81, Fig. 9), and the whole formation was divided into three subordinate units. In its current range, the subformation is composed of alternating limestone and marl members, which enclose thin interbeds of marl, in the first case, and of limestone, in the second. All rocks are gray in color. Small and medium-sized organic remains are irregularly distributed. Ostracodes, brachiopods, and crinoids represent principal groups of fossils, and branching tabulates are characteristic of upper beds of the subformation representing stratotype of the upper Tukal Subhorizon. The range of the latter corresponds to the terminal lower–upper Gorstian and is divided into six local chronozones.

(1) Chronozone $tk_2(1)$. Stratotype of the chronozone is 6.8 m thick and corresponds to beds 3/29, 3/30, and 2/4–2/15. Principal rocks of the unit are greenish gray dolomitic marls with gray lime clots and minor lenses of detrital limestone. These rocks enclose lentils of gray to dark-gray, and platty and foliated detrital clayey limestones. Small, irregularly distributed organic remains occur locally in accumulations. The dominant taxon is *Hyattidina acutisummitatus.*

(2) Chronozone $tk_2(2)$. The stratotype of the chronozone is 6.0 m thick and corresponds to beds 2/16–2/18 and 1/1–1/9. Primary rocks of the chronozone are gray to dark-gray, thin-platty, dolomitic-clayey, detrital limestones displaying even or knobby bedding planes. Subordinate rocks in thin lentils are represented by greenish to dark gray marl with local lime clots, gray dense detrital limestone, and thin stromatoporoid biostromes. Small and medium-sized fossils are distributed irregularly, being concentrated only in detrital limestones. *Sibiricrinus helenae* is the dominant taxon.

(3) Chronozone $tk_2(3)$. Stratotype of the chronozone is 6.7 m thick and corresponds to beds 1/10–1/16, 4/2, and 4/3. The lower beds 1/10–1/14 are composed of gray to greenish friable dolomitic marl enclosing lime clots and thin lentils of gray platty detrital-clayey limestones bearing autoclasts. They show a platty structure and knobby bedding planes. Small and rare fossils are dispersed in marl beds and concentrated in limestones. *Beyrichia quadricornuta* is dominant here. The upper bed sequence (1/15, 1/16, 4/2, and 4/3) includes gray to dark-gray, platty clayey-detrital limestones enclosing laminae of green marl and exposing knobby bedding planes with exsiccation cracks. Small abundant organic remains are buried in the dispersed state or in accumulations. Their dominant taxon is *Murchisonia cingulata.*

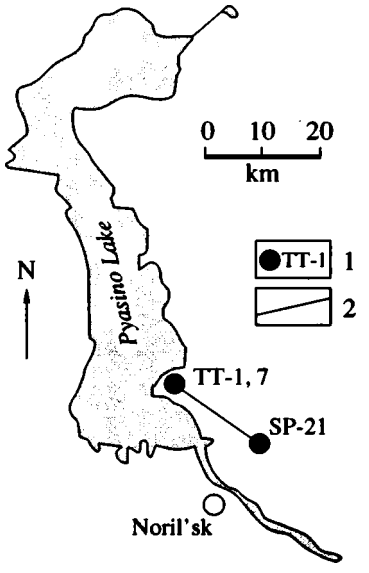

Fig. 9. Borehole localities in the Noril'sk region: (1) borehole indices; (2) profile of the Postnichnyi Formation.

(4) Chronozone $tk_2(4)$. Stratotype of the chronozone is 7.9 m thick and corresponds to beds 4/4–4/12. The lower beds 4/4–4/6 are composed of gray marl enclosing gray fine-grained lime clots and their chains, and also thin lentils of gray clastic-detrital limestones. Small and rare fossils are mostly concentrated in lentils of detrital limestones. *Eukloedenella kureikensis* is dominant here. The upper beds 4/7–4/12 consist of greenish to gray, thin-platty and crumbly conchoidal marls intercalated with large greenish gray lentils of fine-banded thin-platty dolomitic limestones and with thin lenses of detrital limestone and dolomite. Small and rare fossils are concentrated only in detrital limestones. *Leperditia lumaea* is dominant at this level.

(5) Chronozone $tk_2(5)$. Stratotype of the chronozone is 5.1 m thick and corresponds to beds 4/13 and 4/14. The lower bed 4/13 of dark-gray massive polydetrital limestone yields large and medium-sized fossils usually dispersed and locally accumulated. Fossils are often enveloped by mesophytoliths. *Murchisonia cingulata* is dominant in the beds. The upper Bed 4/14 of gray dolomitic marl bears abundant lime nodules of irregular shape, which are light-gray in color and clayey-detrital in composition. Rare organic remains

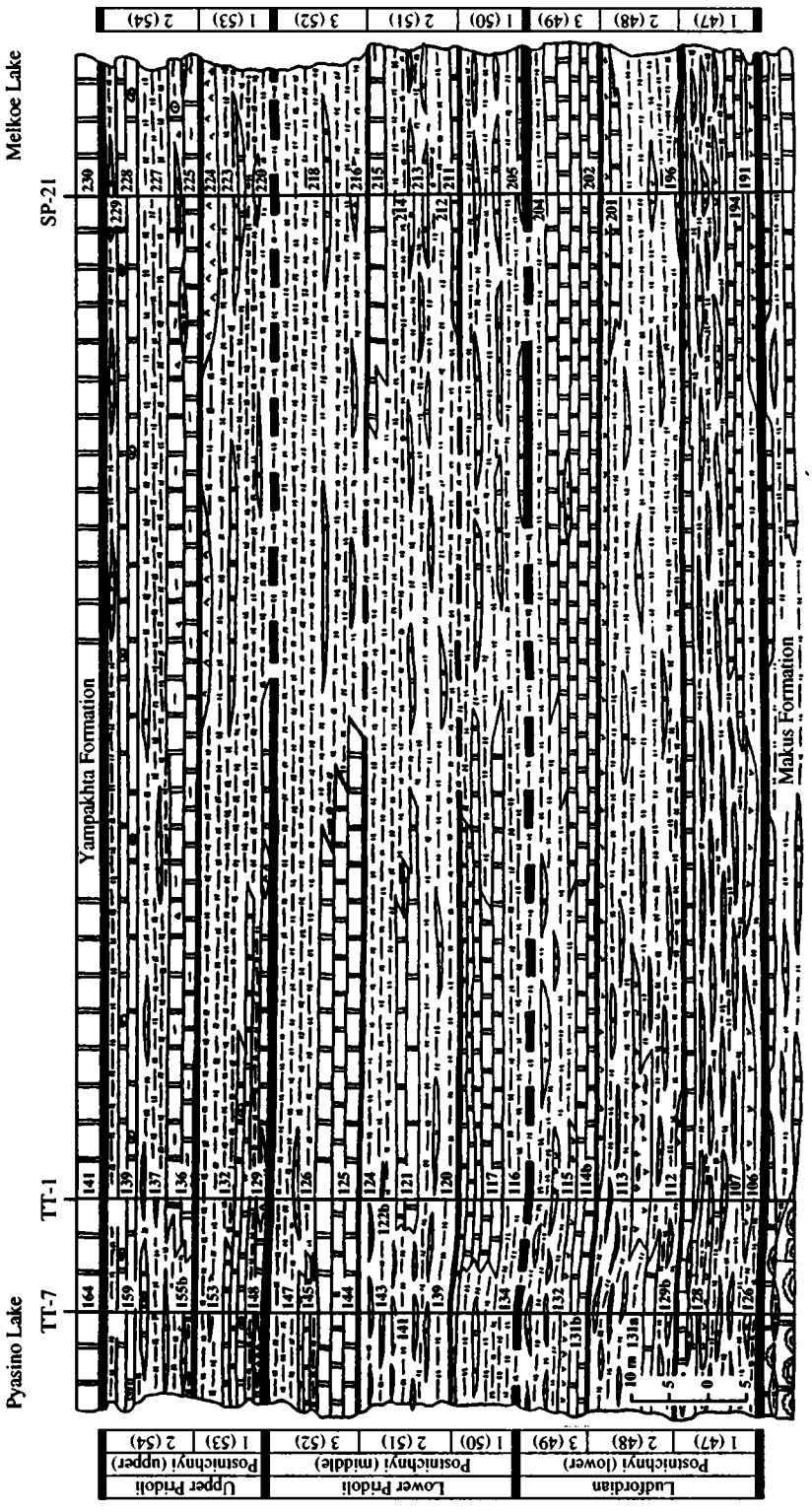

Fig. 10. Stratigraphy, lithology, and characteristic fossil groups of the Postnichnyi Formation (Ludfordian–Pridoli), profile between boreholes TT-7 and SP-21 (symbols as in Fig. 8).

mainly occur in detrital limestones. Their dominant taxon is *Eohowellella minimus*.

(6) Chronozone $tk_2(6)$. Stratotype of the chronozone is 7.7 m thick and corresponds to beds 10/28–10/40. The lower beds 10/28–10/31 are composed of black detrital-nodular bituminous silty limestones showing a vaguely bedded structure and enveloping stromatolitic biostromes and lenticules of detrital limestone. Small and medium-sized fossils are dispersed irregularly, being abundant only at the certain bed levels. The upper part of this bed sequence encloses accumulations of branching tabulates. *Murchisonia cingulata* is the dominant taxon. The middle beds 10/32–10/36 of thin-platty to foliated limestones and dolomites are intercalated with lenticular interbeds of algal stromatolites and organogenic-detrital limestone bearing branching tabulates. Small organic remains are concentrated at particular levels, where *Parastriatopora kureikiana* is dominant. The upper sequence of beds 10/37–10/40 mostly consists of greenish gray dolomites and dark-gray limestones with occasional organic remains. The dominant taxon is *Herrmannina nana*.

The Noril'sk region in the northwest of the Siberian platform (Fig. 1) is the type stratigraphic area of the Postnichnyi Horizon (Ludfordian–Pridoli interval). A profile of exposures characterizing the Postnichnyi Formation is shown in Fig. 9.

The *Postnichnyi Formation* (Fig. 10) was distinguished in 1982 (Predtechenskii and Tesakov, 1982, p. 19) and named after the synonymous creek, the left tributary of the Imangda River. The stratotype of the formation is 81.0 m thick and corresponds to beds 106–140 in the core section of borehole TT-1. Characteristic rocks of the formation are variegated banded marls (green, greenish gray, and dark-gray) intercalated with large lentils of gray platty dolomites and smaller lentils of bluish gray anhydrite and gray brecciated dolomites. The rocks yield scarce organic remains. The formation boundaries are distinct, and its rocks are poorly exposed within the negative relief forms. The sequence in question is divided into three subformations.

The **lower Postnichnyi Subformation** ps_1 was initially defined as the Lower Sequence (Predtechenskii and Tesakov, 1982, p. 19). Stratotype of the subformation is 29.0 m thick and corresponds to beds TT-1/106–TT-1/115 of gray to greenish gray dolomitic marls intercalated with lentils of gray dolomite, anhydrite, and white gypsum. The distinct lower boundary of the unit is placed at the occurrence level of gypsum lentils and banded gypsum-bearing domerites. This is the type unit of the lower Postnichnyi Subhorizon corresponding in stratigraphic range to the Ludfordian Stage. Three local chronozones are distinguished within the unit.

(1) Cronozone $ps_1(1)$. Stratotype of the chronozone is 9.3 m thick and corresponds to beds TT-1/106–TT-1/111 of gray, green, and greenish gray domerites enclosing frequent lentils of gray banded dolomites, dolomitic breccias, and white gypsum.

(2) Cronozone $ps_1(2)$. Stratotype of the chronozone is 11.2 m thick and corresponds to beds TT-1/112–TT-1/114a of gray and greenish gray banded domerites intercalated with rare lentils of massive dolomites and anhydrite-dolomitic breccias.

(3) Cronozone $ps_1(3)$. Stratotype of the chronozone is 7.9 m thick and corresponds to beds TT-1/114b and TT-1/115 displaying the facies variability. In central part of the region (borehole SP-21), the beds consist of dolomite with anhydrite lentils. In other areas, they are represented by domerites enclosing lentils of anhydrite and banded dolomite.

The **middle Postnichnyi Subformation** (ps_2) is described here for the first time. Stratotype of the subformation is 31.5 m thick and corresponds to beds TT-1/116–TT-1/128 consisting of gray to dark-gray banded dolomitic marls, which are intercalated with size-variable lentils of gray banded dolomite and thin rare lenticules of anhydrite. The lower boundary is indistinct. The stratotype exemplifies the middle Postnichnyi Subhorizon, an equivalent of the lower Pridoli, and is divided into three local chronozones.

(1) Chronozone $ps_2(1)$. Stratotype of the chronozone is 8.5 m thick and corresponds to beds TT-1/116–TT-1/119 of gray banded domerites intercalated with rare large lentils of gray banded dolomite and with thin lentils of identical dolomite and gray anhydrite.

(2) Chronozone $ps_2(2)$. Stratotype of the chronozone is 11.8 m thick and corresponds to beds TT-1/120–TT-1/124 of gray domerites intercalated with rare thin and medium-thick lentils of gray banded dolomite and anhydrite.

(3) Chronozone $ps_2(3)$. Stratotype of the chronozone is 11.2 m thick and corresponds to beds TT-1/125–TT-1/128. In the center of the region (borehole SP-21), the beds consist of domerites and span the whole range of the chronozone. In northwestern areas (boreholes TT-1.7), the lower domerite beds contain intercalations of dolomite.

The **upper Postnichnyi Subformation** ps_3 is also described here for the first time. Stratotype of the subformation is 20.5 m thick and corresponds to beds TT-1/129–TT-1/140 of two principal rock types: gray to dark-gray sulfate-bearing domerite, whose beds are intercalated with lentils of gray banded dolomite and light-colored anhydrite; and gray banded dolomite enclosing anhydrite segregations. The stratotype characterizes the upper Postnichnyi Subhorizon identical to the upper Pridoli substage and includes two local chronozones.

(1) Chronozone $ps_3(1)$. Stratotype of the chronozone is 9.1 m thick and corresponds to beds TT-1/129–TT-1/135 of gray to dark-gray banded domerites intercalated with rather large lentils of gray banded dolomite and gray, locally brecciated anhydrite.

(2) Chronozone $ps_3(2)$. Stratotype of the chronozone is 11.2 m thick and corresponds to beds TT-1/136–TT-1/140. Two beds (TT-1/136 and TT-1/139) of gray dolomite include occasional gypsum segregations and authigenic breccia. They yield small and scarce fragments of eurypterids and bivalves. These beds alternate with beds TT-1/137, TT-1/138, and TT-1/140 consisting of gray banded domerites with lenticules of gray dolomite, authigenic breccia, and light-colored anhydrite.

The last beds are overlain by gray massive thick-platty dolomites of the Devonian Yampakhta Formation, which enclose stromatolites (Predtechenskii and Tesakov, 1982).

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Reviewer B.S. Sokolov

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Biozonation of the Syuran Horizon of the Bashkirian Stage in the South Urals as Indicated by Ammonoids, Conodonts, Foraminifers, and Ostracodes

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Abstract—Comprehensive biostratigraphic studies of the Syuran and Bogdanovskii horizons were undertaken in the type area of both horizons in the South Urals. Successions of ammonoid, conodont, foraminiferal and ostracode assemblages were thoroughly studied in the stratotypes of the horizons. A new refined correlation of the ammonoid *Homoceras-Hudsonoceras* and *Reticuloceras-Bashkortoceras* genus-zones to coeval zones of foraminifers, conodonts, and ostracodes in the South Urals and other regions is suggested.

Key words: *Bashkirian Stage, Bogdanovskii Horizon, Syuran Horizon, stratotype, ammonoids, conodonts, foraminifers, ostracodes, biostratigraphy.*

INTRODUCTION

In connection with the elaboration of the global Carboniferous scale, the detailed characterization of its units, the substantiation of their boundaries, and the precise determination of stratigraphic ranges are the most important objectives. The Bashkirian Stage, which was recognized by S.V. Semikhatova in the mountains of Bashkortostan in 1934, is the lower unit of the Middle Carboniferous in the North Eurasian stratigraphic scale and, undoubtedly, can be used worldwide. The lower boundary of the unit coincides with the Mid-Carboniferous boundary, the GSSP of which was chosen in the Arrow Canyon section, Nevada, USA (Lane *et al.*, 1994). The boundary is drawn at the bases of the ammonoid *Homoceras-Hudsonoceras* Genus-zone and the conodont *Declinognathodus noduliferus* Zone. Since 1968, stratigraphic scales of the Bashkirian Stage begin in the Urals with the Syuran Horizon. The attributive "Syuran" was first used by Librovich (1947, p. 59) to describe the limestone facies at the Suren' (Syuran) River in the western slope of the South Urals. The facies is characterized by the coexistence of goniatite and foraminifer faunas. The Bol'shaya Suren' River section near the Bogdanovskii Farmstead (now the Novyi Bogdanovskii Settlement) was proposed by Librovich and Sultanaev (1977, p. 362) as the stratotype of the Syuran Horizon and became a site of long-term investigations. Ammonoids of the section were first described by Krestovnikov (1935) and later studied by Librovich (1947), Ruzhentsev and Bogoslovskaya (1978). Along the left tributary of the Bol'shaya Suren' River, 1 km to the south of the Bogdanovka Village (now the Novo-Bogdanovskii Settlement), Einor *et al.* (1973a, 1973b) recognized the lower part of the Syuran beds with ammonoids of the *Homoceras* Zone as the independent Bogdanovskii Horizon, which they studied in two sections. The stratotype of the horizon was established in the Novyi Bogdanovskii Settlement section (Exposure 4), the upper part of which corresponds, according to the cited authors, to the ammonoid zone H_2 (Fig. 1). The lower part of the horizon is exposed 2 km to the east, in the section near the Kugarchi Village. Reitlinger (1980) described foraminifers from the Bogdanovskii Horizon. Later, Einor (1992) found here some ammonoids characteristic of zone H_1. However, his unique section, where ammonoids are associated with conodonts, foraminifers, and ostracodes at several stratigraphic levels, has never been adequately studied. The differently interpreted stratigraphic range of the Syuran Horizon was a subject of long-lasting discussions (Nikolaev, 1989; Einor, 1992). For example, in the early stratigraphic schemes for the Urals (*Unifitsirovannye...*, 1968, 1980), the horizon corresponds to the *Reticuloceras* Zone, whereas, according to the latest scheme (*Stratigraphicheskie...*, 1993), it spans the interval of two *Homoceras-Hudsonoceras* and *Reticuloceras-Bashkortoceras* genus-zones. In this work, we use the latter one. We studied and described in detail the section, which can be proposed as a stratotype of both the Syuran and Bogdanovskii horizons. We also studied the bed-by-bed collections of ammonoids, conodonts, foraminifers, and ostracodes.

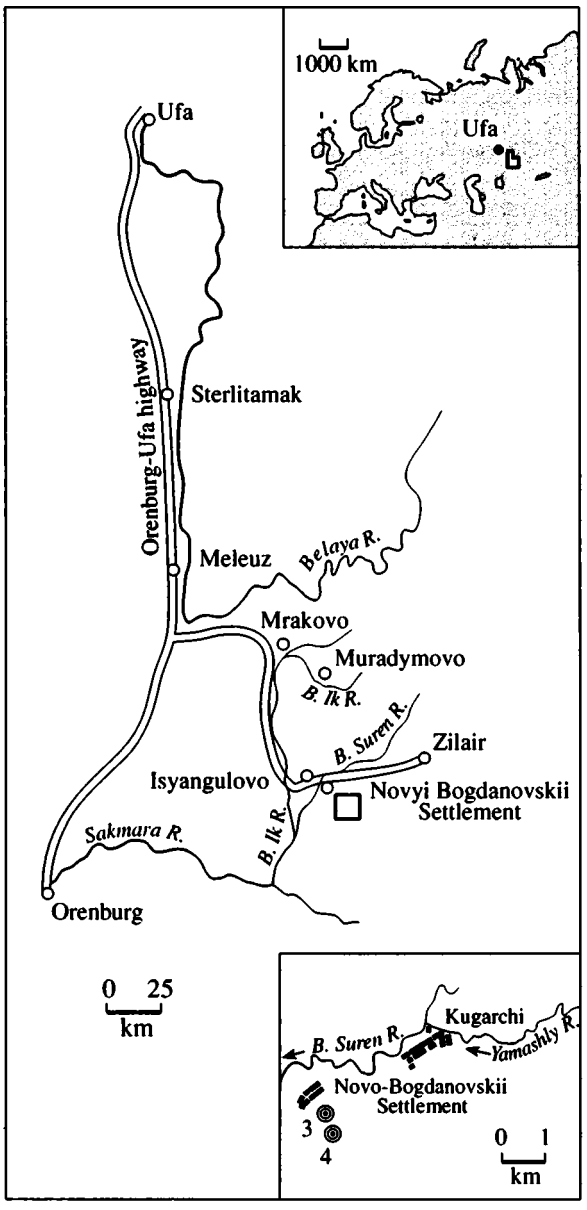


Fig. 1. Locality of the Bogdanovskii Horizon stratotype in Bashkortostan.

MATERIAL

In 1986–1997, we investigated the Syuran and Bogdanovskii type sections using the bed-by-bed method. To establish the ammonoid succession, we revised their specimens collected by Einor, Furdui and Aleksandrov in 1971 from the same exposures and deposited at the Paleontological Institute of Russian Academy of Sciences. The studied ammonoids are stored at this Institute as collection no. 4715, whereas the collections of conodonts, foraminifers and ostracodes are stored under nos. 104, 121, and 66 at the Institute of Geology of the Ufa Scientific Center, Russian Academy of Sciences.

DESCRIPTION OF THE SECTION

The section is located on the western slope of the South Urals in the Zilair subzone of the Central Uralian structural-facial zone (*Stratigraficheskie...*, 1993), 1 km to the southeast of the Novo-Bogdanovskii Settlement at the left bank of the Bol'shaya Suren River (Fig. 1). The meridional anticlinal structure (the Bogdanovskii anticline) can be traced here in two main exposures (Fig. 2). The anticline is composed of thin- to medium-bedded, locally foliated carbonate and terrigenous-carbonate deposits. In our work, we followed the exposure numeration after Einor and co-authors (Einor *et al.*, 1973a; 1973b; Einor, 1992). Exposure 3 is

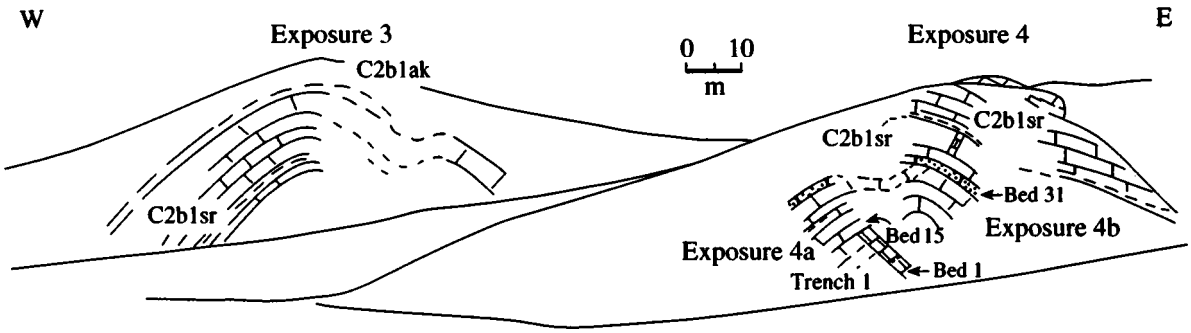


Fig. 2. Schematic profile of the Mt. Kamennaya demonstrating relations between exposures 3 and 4 (a and b).

nearest to the Novyi Bogdanovskii Settlement, and Exposure 4 is located 200 m to the south of it.

Exposure 4 reveals the oldest beds.

EXPOSURE 4

The exposure is located on the southeastern slope of the Mt. Kamennaya, where the beds yield ammonoids of the *Homoceras-Hudsonoceras* Genus-zone. The exposure can be suggested as the stratotype of the Bogdanovskii Horizon. It consists of two parts (4a and 4b) situated in the western and eastern limbs of the anticlinal structure, respectively, being separated by a 20-m-wide unexposed interval (Fig. 2). The composite column of the exposure was presented in the previous publication (Kulagina *et al.*, 1992); numbers of beds in the present description are the same.

Western Part of Exposure 4 (Exposure 4a)

Exposure 4a in the western limb of the anticlinal fold, where attitude of limestone beds corresponds to azimuth 270–295° and dip angle 25–30°, is studied in the recent prospecting trench, where neither Librovich nor Einor were able to observe the lowermost beds 1–13 (Fig. 3). At the base, there are yellow clays (0.05 m) resting on almost black cherts. The following succession of members and beds can be observed here (from the base upward).

Member 1:

(1) gray organogenic clastic limestone (litho-bioclastic grainstone) siliceous in the lower part and bearing bryozoans, algae, crinoids, foraminifers, and ostracodes (0.45 m thick, samples P-05B, P-05A, P-05);

(2) yellow clay with fragments of siliceous limestones and cherts (0.05 m thick);

(3) gray detrital limestone (grainstone) with siliceous-clayey interbeds at the base, which contain crinoids, brachiopods, ammonoid larvae, and sponge spicules (0.25 m thick, Sample P-1B);

(4) gray organogenic-detrital limestone with pelitomorphic fine-grained cement (wackestone) bearing small brachiopods, foraminifers, and ostracodes. The thickness is 0.65 m (Sample P-1A);

(5) gray organogenic-detrital limestone with the clastic upper part (wackestones to litho-bioclastic grainstones) containing foraminifers, algae, fragments of brachiopods, bryozoans, echinoderms, corals, clastic material (lithoclasts and sandy grains), and rare ooliths (0.83 m thick, samples P-2, P-3ab, 04).

Member 2:

(6) gray, slightly clayey micritic limestone (micrite) with rare unilocular foraminifers and spheres (0.35 m thick, Sample P-4).

Member 3:

(7) gray organogenic-clastic limestone (litho-bioclastic grainstone and packstone) grading into calcareous breccia and bearing ooliths, foraminifers, bryozoans, crinoids, and brachiopod fragments (0.15 m thick, Sample 03/2);

(8) spicule-bearing gray micritic clayey to silty wackestone (0.1 m thick, Sample P-5);

(9) gray fine-grained to clotted micritic limestone (mudstone-wackestone) with foraminifers, ammonoids, and sponge spicules (0.57 m thick, Sample P-6, P-7);

(10) gray organogenic-clastic limestone with pelitomorphic fine-grained matrix (bioclastic wackestone-packstone); the bed yields ooliths, oncolites, foraminifers, and ammonoids (0.3 m thick, Sample 03);

(11) gray, slightly clayey micritic limestone (mudstone) with rare fine detritus (0.85 m thick, Sample 02).

Member 4:

(12) gray spicule-bearing clayey micritic limestone (0.2 m thick, Sample P-8);

(13) gray clotted micritic limestone (1 m thick, Sample 01).

Member 5:

(14) gray clotted micritic limestone with spheres (biopelsparite) and small foraminiferal shells (0.85 m thick, Sample P-9);

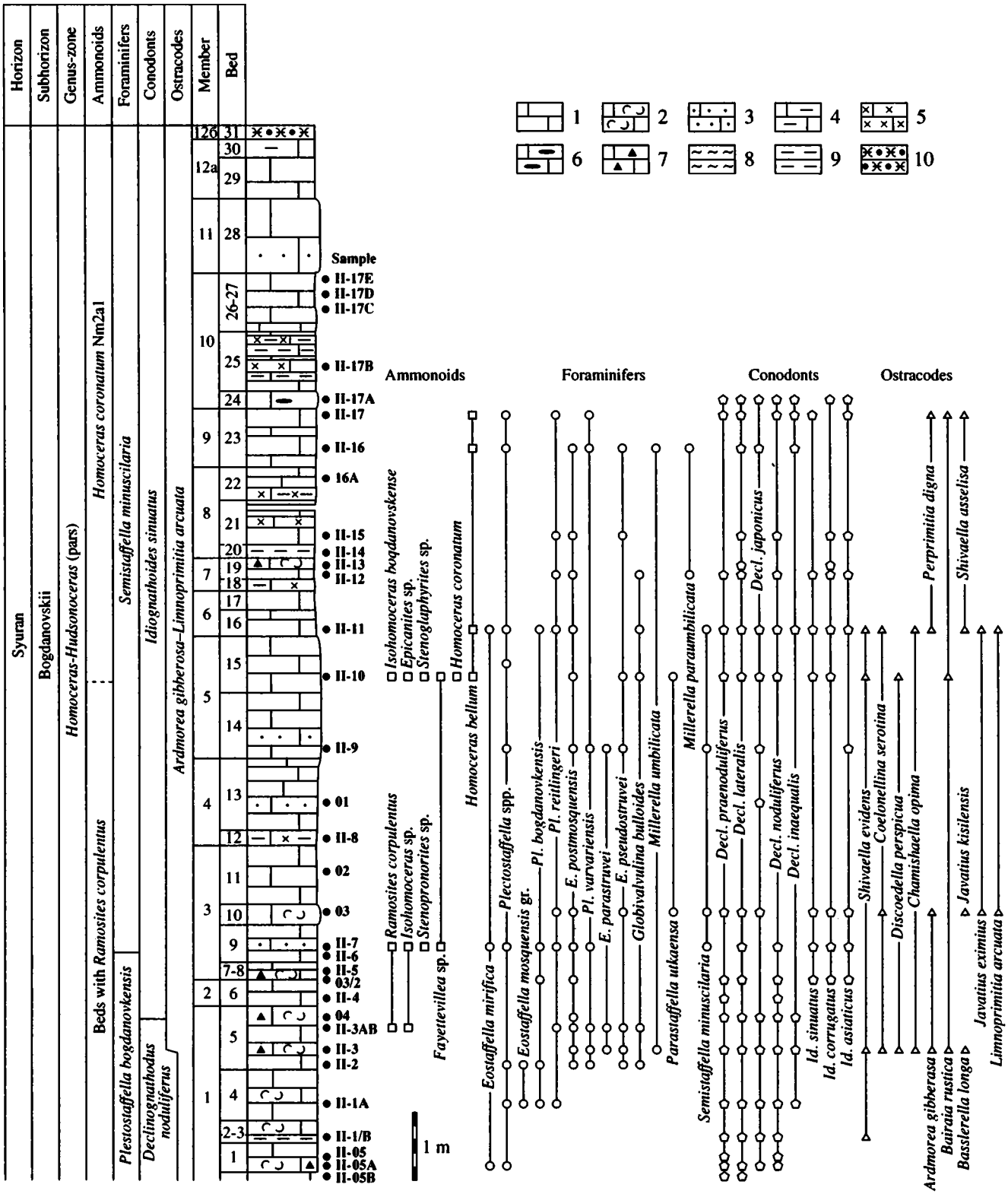


Fig. 3. Middle Carboniferous section of the Mt. Kamennaya (Exposure 4a) and distribution of ammonoids, foraminifers, and conodonts: (1) micritic limestone or organogenic wackestone; (2) organogenic and bioclastic limestone or packstone and greenstone; (3) clotted limestone or pelmicrite and pelsparite); (4) clayey limestone; (5) spicule-bearing limestone; (6) limestone with chert concretions; (7) clastic (bio-lithoclastic) limestone; (8) clayey-siliceous and siliceous shales; (9) mudstone; (10) sandstone.

(15) dark gray micritic limestone (wackestone) with accumulations of foraminifers and ammonoids (0.65 m thick, Sample P-10).

Member 6:

(16) gray micritic limestone with sporadic organogenic interbeds and insignificant sandy admixture; the bed (bioclastic wackestone) bears frequent foraminifers, ostracodes, and fragments of brachiopods and ammonoids (0.5 m thick, Sample P-11);

(17) gray micritic limestone (micrite) 0.42 m thick.

Member 7:

(18) gray slightly clayey micritic limestone with sponge spicules (0.2 m thick);

(19) gray fine-grained spherical-clotted micritic limestone; the upper part of the bed includes an interbed of organogenic-clastic limestone (litho-bioclastic packstone) with algae, crinoids, brachiopods, and ammonoids (0.4 m thick, samples P-12, P-13).

Member 8:

(20) limy mudstone with rare sponge spicules and a silty admixture (0.2 m thick, Sample P-14);

(21) gray micritic limestone (micrite) and clotted limestone (biopelmicrite) with a thin (0.02 m) interbed of spicule-bearing limestone (1.4 m thick, Sample P-15);

(22) gray micritic limestone (micrite) with accumulations of foraminifers, ammonoids, ostracodes, brachiopods, and bryozoan fragments (wackestone); at the base of the bed, there is an interbed (0.02 m) of brownish gray clayey spicule-bearing limestone (0.65 m thick, Sample 16a).

Member 9:

(23) gray micritic limestone, the upper part of which is organogenic and encloses foraminifers and ammonoids (0.9 m thick, samples P-16, P-17).

Member 10:

(24) gray fine-grained ammonoid-bearing limestone with thin chert lenses (0.3 m thick, Sample P-17A);

(25) interlaminated micritic, clayey, and spicule-bearing limestones (0.8 m thick, Sample 17B);

(26–27) gray micritic limestone (1.2 m thick, Sample P-17C, P-17D, P-17E).

Member 11:

(28) micritic and clotted limestone with detrital accumulations; along the strike, the bed is divided into two parts visible in places at the slope further toward Exposure 4b, where it forms at first synclinal, and then anticlinal folds (1.3 m thick, samples 174, 172, 173 in Exposure 4b).

Member 12a:

(29) gray micritic limestone 0.65 m thick;

(30) gray clayey micritic limestone 0.35 m thick.

Member 12b:

(31) brownish gray fine-grained microlaminated sandstone (0.15–0.2 m thick, Sample 28A from Exposure 4b).

Eastern Part of Exposure 4 (Exposure 4b)

The exposure is located in the right limb of the anticlinal fold, where the attitude of the beds corresponds to azimuth 70–95° and dip angle 30–40° (Fig. 4). The exposure includes members 5–12b of the Exposure 4a. The higher stratigraphic levels are divided into successive members 12–20.

Member 12b:

(32) gray micritic limestone with sponge spicules (0.15 m thick, Sample P-28);

(33) brownish gray fine- to medium-grained sandstone 0.4 m thick;

(34) gray micritic limestone with clayey interbeds and sponge spicules (0.6 m thick, Sample P-29).

Member 13:

(35) gray micritic limestone 1.1 m thick.

Member 14:

(36) organogenic-clastic limestone grading into calcareous breccia in the lower part and micritic limestone in the upper part (0.85 m thick, samples P-30, P-30A, P-31).

Members 15–17 and beds 37–40 are exposed in the prospecting trench.

Member 15:

(37) fine intercalation of micritic and clayey limestones and cherts (0.9 m thick, Sample P-32);

(38) gray micritic limestone (1.7 m thick, Sample P-33A).

Member 16:

(39) gray micritic limestone, an upper part of the bed (0.5 m) is seen in the natural exposure (1.8 m thick, Sample P-35).

Member 17:

(40) gray micritic limestone with clayey interbeds (2.7 m thick, Sample 36).

Member 18:

(41) gray micritic limestone (1.7 m thick, Sample 37);

(42) Gray spicule-bearing micritic limestone with clayey interbeds (3 m thick, samples 38, 39);

(43–44) gray micritic limestone with clayey interbeds and rare sponge spicules (4.9 m thick, Sample 40).

Member 19:

(45) micritic limestone (0.7 m) at the base and frequently alternating micritic and clayey limestones above (3.8 m thick, samples 41, 41A, 42);

(46) gray micritic fine-grained limestone, locally pachyspherical-clotted and bearing rare foraminifers and organic detritus (4.8 m thick, samples 43, 44).

Member 20:

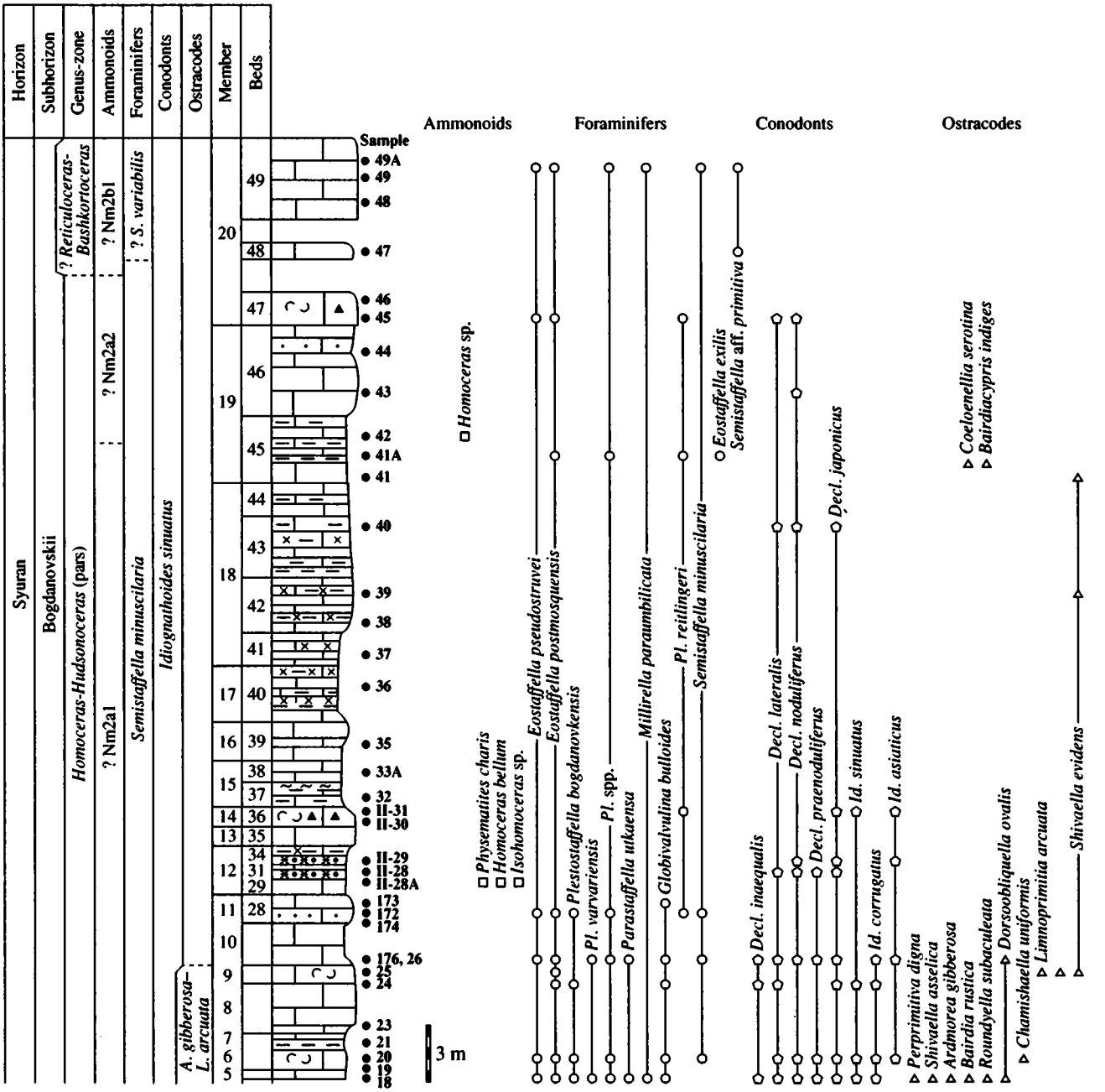


Fig. 4. Middle Carboniferous section of the Kamennaya Mountain (Exposure 4b) and distribution of ammonoids, foraminifers, and conodonts (symbols as in Fig. 3).

(47) gray organogenic-detrital limestone with pelitomorphic fine-grained matrix; the bed yielding foraminifers, bryozoans, and echinoderms forms synclinal and anticlinal folds traceable along the slope (1 m thick, samples 45, 46).

The unexposed interval 1.4 m wide.

(48) gray micritic clayey limestone (0.2 m thick, Sample 47).

The unexposed interval 1.2 m wide.

(49) gray micritic limestone with rare interbeds of organogenic-detrital limestone and fine-grained matrix (4 m thick, samples 48, 49, 49A, 50). The bed is destroyed during work in the quarry.

EXPOSURE 3

The exposure is located 200 m to the northwest of Exposure 4 on the southern slope of a mountain near Mt. Kamennaya (Fig. 2). Deposits of the site bear

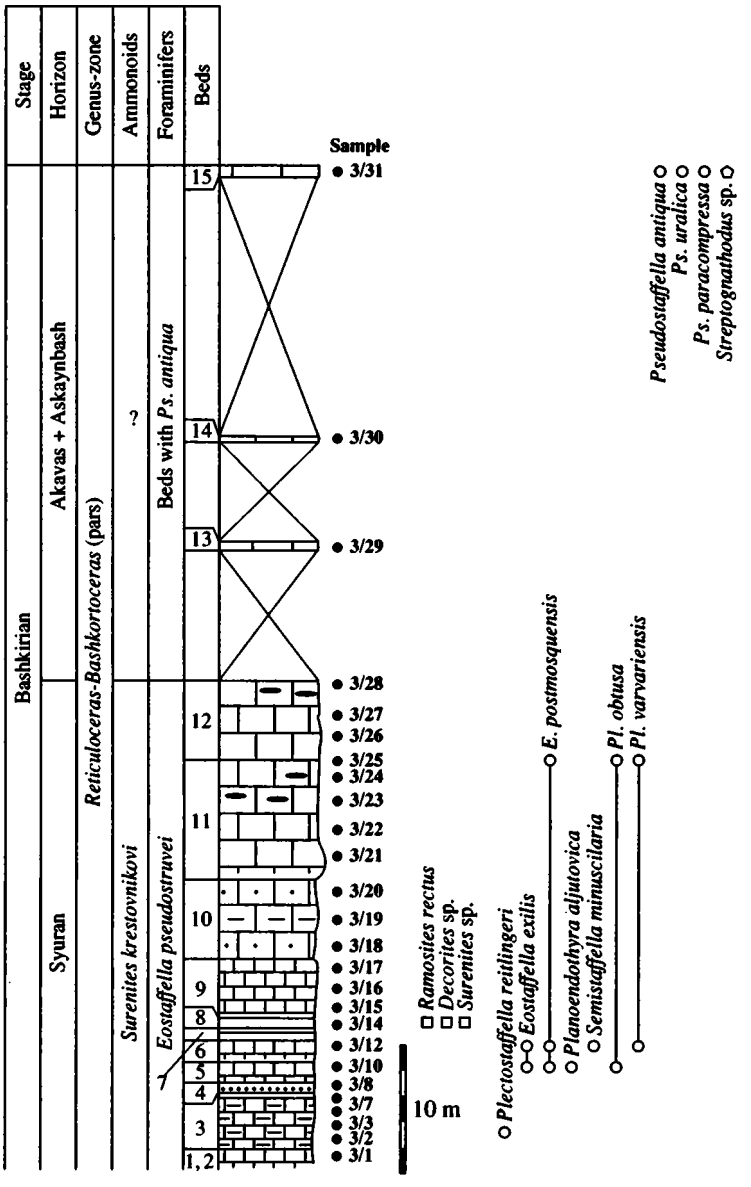


Fig. 5. Middle Carboniferous section of the Mt. Kamennaya (Exposure 3) and distribution of ammonoids, foraminifers, and conodonts (symbols as in Fig. 3).

ammonoids characteristic of the *Reticuloceras-Bashkortoceras* Genus-zone (Fig. 5). In the western wing of the fold, the following beds are exposed (from the base upward).

(1) Gray micritic limestone (0.25 m thick, Sample 3/1).

(2) Gray micritic limestone enclosing interbeds with small foraminiferal shells and accumulations of fine detritus along bedding planes (1.3 m thick, Sample 3/2).

(3) Gray micritic, silty-clayey limestone with sponge spicules, fine-grained interbeds, and local carbonaceous and clayey microlaminae (3.8 m thick, samples 3/3–3/7).

(4) Organogenic calcarenite (0.3 m thick, Sample 3/8).

(5) Gray spicule-bearing micritic and weakly clayey limestone with an interbed of fine-grained micritic limestone bearing foraminifers and sponge spicules (1.8 m thick, samples 3/9, 3/10).

(6) Gray micritic, clayey and microlaminated limestone with sponge spicules accumulated along bedding planes (1.4 m thick, samples 3/11, 3/12).

(7) Gray organogenic-detrital limestone with pelitomorphic fine-grained matrix; the bed yields foraminifers, ammonoids, and sponge spicules (0.2 m thick, Sample 3/13).

(8) Gray organogenic-clastic limestone with micritic matrix (bio-lithoclastic wackestone-pack-

stone) consisting of partly rounded large tests of foraminifers *Tolypammina*, ammonoids, detritus, ooliths, pellets, and intraclasts (0.25 m thick, Sample 3/14).

(9) Gray micritic clayey limestone with sponge spicules (3.6 m thick, samples 3/15–3/17).

(10) Alternating micritic and clayey spicule-bearing limestones (8 m thick, samples 3/18, 3/20).

(11) Gray micritic limestone with chert nodules and brownish gray clayey spicule-bearing interbeds (9.1 m thick, samples 3/21, 3/22, 3/23, 3/24, 3/25).

(12) Gray micritic limestone (6 m thick, samples 3/26, 3/27, 3/28).

The unexposed interval 10 m wide.

(13) Gray micritic limestone (0.2 m thick, Sample 3/29).

The unexposed interval 7 m wide.

(14) Gray micritic limestone with small cherty concretions (0.2 m thick, Sample 30).

The unexposed interval 19 m wide.

(15) Gray organogenic-clastic limestone (bioclastic wackestone) with abundant foraminifers, algae, and bryozoans (0.3 m thick, Sample 3/31).

BIOSTRATIGRAPHY

The ammonoid, conodont, foraminifer, and ostracode zonations were recognized in the exposures studied. They were correlated with those characterizing the boundary intervals between the Upper and Middle Carboniferous in other sections of the Urals and Central Asia, and also between the Lower and Upper Carboniferous in Western Europe and between Mississippi and Pennsylvania in North America.

AMMONOIDS

Ammonoids were found in exposures 4a (six levels), 4b (two levels), and 3 (one levels). The ammonoids from Exposure 3 are similar to the assemblage described by Ruzhentsev and Bogoslovskaya (1978, p. 15) from their Sample 3. Other samples studied by Ruzhentsev and Bogoslovskaya were collected in the Bogdanovskii section to the east of Sample 3 in the same exposure. Analysis of ammonoids from nine levels allowed us to establish six successive assemblages typical of two ammonoid *Homoceras-Hudsonoceras* and *Reticuloceras-Bashkortoceras* genus-zones (Plate I).

The first genus-zone includes ammonoids from beds 5, 9, 15, 16, 23, and 24 of the Exposure 4a and from Bed 31 of the Exposure 4b. The assemblage is similar to that of the *Homoceras coronatum* Zone (the lower part of the *Homoceras-Hudsonoceras* Genus-zone), which was first recognized in Sholak-Sai (Ruzhentsev and Bogoslovskaya, 1978). The 1.8 m-thick limestone (Sample 3a, 3b) below Bed 5 is lacking in ammonoids, and we conventionally referred this interval to the *Homoceras-Hudsonoceras* Genus-zone. In general, the assemblage is very similar to that from the Sholak-Sai and includes the same or similar species. In the generic composition, it appears to be more similar to the coeval highly diverse fauna of Central Asia (Nikolaeva, 1994b, 1995) rather than to the assemblage of Western Europe (Bisat, 1924; Riley, 1987) including only genera *Isohomoceras* and *Homoceras*. In Central Asia, the ammonoid fauna is characterized by the coexistence of the genus *Fayettevillea*, typical of the Early Carboniferous and genus *Isohomoceras* characteristic of the Middle Carboniferous. It is noteworthy that the genus *Proshumardites* is found, together with the genus *Isohomoceras*, in the Sholak-Sai and localities of Central Asia.

The *Reticuloceras-Bashkortoceras* Genus-zone comprises beds of the Exposure 3. The assemblage includes ammonoids *Surenites krestovnikovi, Decorites* sp. and *Ramosamples rectus* characteristic of the *Surenites krestovnikovi* Zone (Ruzhentsev and Bogoslovskaya, 1978).

Homoceras-Hudsonoceras Genus-Zone

(1) *Ramosites corpulentus* Beds. Assemblage 1 characterizes beds 5 and 9 located approximately 1.8 m (Sample 3A and 3B) and 3.4 m (Sample 7) above the section base. The assemblage contains representatives of genera *Stenopronorites, Isohomoceras, Fayettevillea*, and *Ramosites*, about 50 specimens altogether.

The assemblage is dominated by *Ramosites corpulentus* Ruzh. et Bogosl. (about 40 specimens), which marks together with *Isohomoceras* forms the lower beds of the *Homoceras-Hudsonoceras* Genus-zone. As the genus *Fayettevillea* is present here, whereas the genus *Homoceras* is missing, this level presumably corresponds to the 0.75-m-thick interval of barren limestones between the latest ammonoids of the *Fayettevillea-Delepinoceras* Zone (Sample 3, the *Delepinoceras bressoni* Zone) and the earliest ammonoids of the *Homoceras-Hudsonoceras* Genus-zone (Sample 4, the *Homoceras coronatum* Zone) in the Sholak-Sai section (Ruzhentsev and Bogoslovskaya, 1978). Assemblage 1, in contrast to that of the Sholak-Sai section, does not contain *Homoceras coronatum*. Therefore, the enclosing deposits are referred to as the *Ramosites corpulentus* Beds. The assemblage is supposed to be coeval with that of the *Isohomoceras ventrosum* Zone of Central Asia (Nikolaeva, 1994b) and younger than that of the *Isohomoceras subglobosum* Zone of Central Asia and England (Riley *et al.*, 1987; Nikolaeva, 1994a, 1994b).

The *Ramosites corpulentus* Beds are 4.06 m thick in Exposure 4a.

(2) *Homoceras coronatum* Beds. Assemblage 2 was found at two levels: one corresponding to beds 15 and 16 (samples 10 and 11) and another located 6.6 to 7.25 m above the section base. Einor *et al.* (1973a)

reported on ammonoids of the genus *Homoceras* occurring, as we think, at the same levels.

The assemblage includes *Epicanites* sp., *Stenoglaphyrites* sp., *Isohomoceras bogdanovkense* Nikolaeva, *Homoceras bellum* Nikolaeva, *H. coronatum* (Haug), and *Fayettevillea* sp. Coexisting genera *Homoceras* and *Isohomoceras* were first recorded in the Sholak-Sai (Ruzhentsev and Bogoslovskaya, 1978), but this atypical of many sections in Europe and Asia (Riley *et al.*, 1987; Nikolaeva and Nigmadganov, 1992; Nikolaeva, 1994a, 1994b). Manger *et al.* (1985) reported on association of these genera in basal beds of the *Homoceras* Zone in Algeria, but the Algerian assemblage may appear to be younger than the described one. The presence of the genus *Homoceras* evidences that Assemblage 2 is younger than the assemblages of the *Isohomoceras subglobosum* and *Isohomoceras ventrosum* zones described by Nikolaeva (1994b) from sections 1–4 at the left bank of the Aksu River and from Section 13 at the Peshkaut Creek of the Tien Shan; the same is correct with respect to the *Isohomoceras subglobosum* Zone of the Stonehead Beck section in England (Riley *et al.*, 1987).

Due to the occurrence of the genus *Fayettevillea* (Plate I, Fig. 9) in Assemblage 2, the enclosing beds of the Bogdanovskii section may be considered somewhat older than the level of Sample 4 in the Sholak-Sai section, in which this genus is missing

Assemblage 3 was found at two levels of Bed 23 (samples P-16, P-17) approximately 10.9 and 11.4 m above the section base. The assemblage includes *Homoceras bellum* Nikolaeva (Nikolaeva, 1999), which was earlier identified as *Homoceras* aff. *haugi* Ruzh. et Bogosl. (Kulagina *et al.*, 1992). This form, very much resembling *Homoceras coronatum*, suggests that the enclosing bed belongs to the *Homoceras coronatum* Zone. In Exposure 4a, the zone is slightly more than 10 m thick.

Assemblages 4 and 5 were found in Exposure 4b.

Assemblage 4 is recovered from Bed 32 (Sample P-28) 11.5 m above the exposure base. it includes *Homoceras bellum, Isohomoceras* sp., and *Physematites charis* Ruzh. et Bogosl. (Plate I). This assemblage is attributable to the *H. coronatum* Zone because the combination of *Homoceras* and *Physematites charis* was found in this zone of the Sholak-Sai section (Ruzhentsev and Bogoslovskaya, 1978).

In Exposure 4b, the beds referred to the *Homoceras coronatum* Zone are approximately 11 m thick. The upper zone boundary has not been established. The overall thickness of deposits spanning the *Homoceras coronatum* Zone is about 33 m.

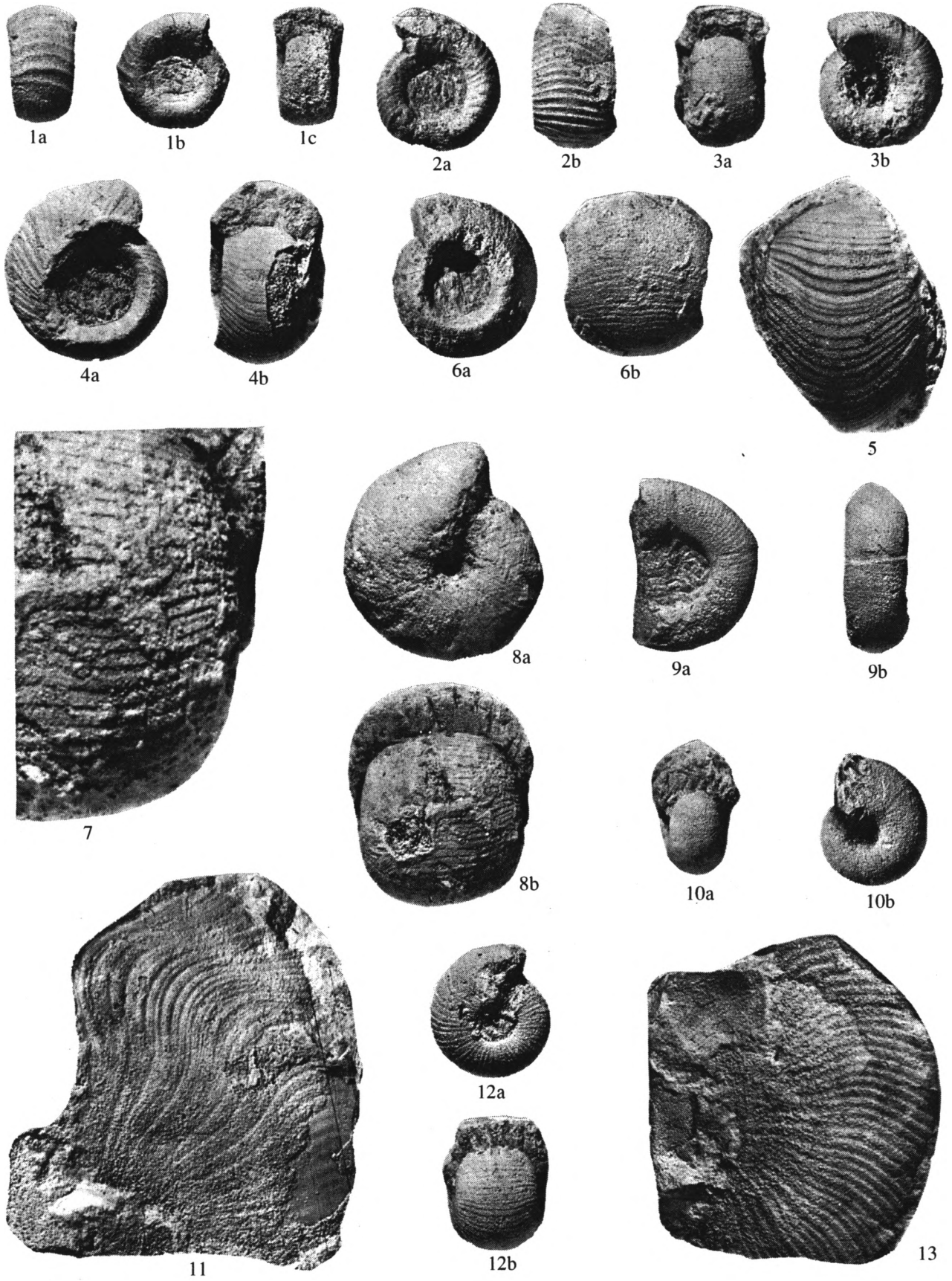
(3) *Homoceras* sp. Beds yield the Assemblage 5. Poorly preserved shells of *Homoceras* sp. were found 1 m higher than Sample 42 (Bed 45), i.e., at the level of 35 m above the section base. The fine sculpture makes these representatives of the genus *Homoceras* similar to the species of the Western European Zone H2 (for example, to *Homoceras undulatum* (Brown). However, the preservation degree is unfavorable for reliable species identification, and we conventionally attributed the beds with this form to the upper part of the *Homoceras-Hudsonoceras* Genus-zone (probably, the *Hudsonoceras proteum* Zone). Thus, we have revised the previous age determination for the 34-m-thick members 13–19 (beds 35–46) overlying Member 12, the sandstone bed of which was considered earlier by Einor *et al.* (1973a) as bearing ammonoids of the *Reticuloceras-Bashkortoceras* Genus-zone. They also considered Member 12 as the uppermost marker of the Bogdanovskii Horizon. The *Homoceras* sp. Beds can correspond to Zone H_2, and the overall thickness of the *Homoceras-Hudsonoceras* Genus-zone is more than 44 m.

Reticuloceras-Bashkortoceras Genus-zone, the Surenites krestovnikovi Zone

The boundary between the *Homoceras-Hudsonoceras* and *Reticuloceras-Bashkortoceras* genus-zones is supposed to be recognizable in Exposure 4b, because ammonoids characteristic of the *Reticuloceras-Bashkortoceras* Genus-zone, which are stored at the Paleontological Institute, originate from this exposure (Einor, 1992; Einor *et al.* 1973b). However, at the top of the mountain, the upper part of the section is destroyed by recent quarry works that are now inaccessible. Assemblage 6 from Bed 8 of Exposure 3 includes *Ramosites rectus* Ruzh. et Bogosl., *Decorites* sp., and *Surenites* sp. and characterizes the lower part of the *Reticuloceras-Bashkortoceras* Genus-zone. Ruzhentsev and Bogoslovskaya (1978) reported on *Surenites krestovnikovi* Ruzh. et Bogosl. from the same bed (Sample 3). According to Ruzhentsev and Bogoslovskaya (1978, p. 15), the beds lying above bear *Baschkirites planus* Ruzh. et Bogosl., *Proshumardites karpinskii* Raus., *Paraschartymites repens* Ruzh. et Bogosl., *Fallacites* aff. *portentosus* Ruzh. et Bogosl., *Phillipsoceras gamarhipaeum* Ruzh. et Bogosl., *P. carinatum* Ruzh. et

Plate I. Ammonoids from the right bank of the Bol'shaya Suren' River, site near the Novyi Bogdanovskii Settlement.

(1–5) *Homoceras bellum* Nikolaeva: (1) Specimen 4475/15, ×3; (2) Holotype 4475/18, ×3; (3) Specimen 4475/30, ×5; (4) Specimen 4475/13, ×3; (5) Specimen 4475/14, ×3; Exposure 4a, (1 and 2) Sample 4/1, (3) Sample P-28, (4 and 5) Sample P-16; (6) *Homoceras coronatum* (Haug), Specimen 4475/20, ×3; Exposure 4a, Sample 4/1 (P-10); (7–8) *Physematites charis* Ruzhencev et Bogoslovskaya, Specimen 4475/40b, ×10 for (7) and ×5 for (8), Exposure 4b, Sample P-28; (9) *Stenoglaphyrites* sp., Specimen 4475/16, ×3, Exposure 4a, Sample 4/1 (P-10); (10) *Fayettevillea* sp., Specimen 4475/17, ×3, Exposure 4a, Sample 4/1 (P-10); (11) *Isohomoceras* sp., Specimen 4475/41, ×3, Exposure 4a, Sample P-7; (12) *Isohomoceras bogdanovskense* Nikolaeva, Holotype 4475/19, ×3, Exposure 4a, Sample 4/1 (P-10); (13) *Ramosites corpulentus*, Specimen 4475/35, ×3; Exposure 4a, Sample P-7.


1a
1b
1c
2a
2b
3a
3b
4a
4b
6a
6b
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7
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9a
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8b
10a
10b
11
12a
12b
13

Bogosl., and *Reticuloceras excultum* Ruzh. et Bogosl.. This assemblage characterizes the upper part of the *Reticuloceras-Bashkortoceras* Genus-zone and can be correlated with assemblages from the *Reticuloceras reticulatum* Zone and, partly, from the *Phillipsoceras gracile* Zone of Western Europe.

CONODONTS

Conodonts of the Bogdanovskii and Syuran horizons of the Suren' River basin were first detected by Furdui (Einor *et al.*, 1973a). In the cited and later publications, Furdui summarized data on conodont assemblages from the Bogdanovka, Yamashly, Uskalyk, and Askyn sections, but without their zonation (Furdui, 1975). The assemblages include forms characteristic of the Serpukhovian and Bashkirian stages.

The conodont collection studied in this work is from Exposure 4, including its oldest, formerly unexposed interval. The collection includes over 1500 specimens from 33 samples. More than 90% of the specimens are represented by platform elements, and about 8% of them correspond to conodonts redeposited mainly from the Serpukhovian, and to a lesser degree, from Famennian and Visean strata. The Bogdanovskii Horizon of Exposure 4 yielded conodonts of the *D. noduliferus* and *Id. sinuatus* zones (Plate II). The first unit is represented by the upper *Late D. noduliferus* Subzone (Pazukhin, 1995). The lower *Early D. noduliferus* Subzone is unexposed. The first occurrence level of *Declinognathodus noduliferus* has not been identified in the type area of the Bogdanovskii and Syuran horizons.

Declinognathodus noduliferus Zone

Late Declinognathodus noduliferus Subzone

The *Late Declinognathodus noduliferus* Subzone spans beds 1–5 of Exposure 4a (thickness 2.6 m). Species of stratigraphic significance are *Declinognathodus inaequalis* (Hig.), *D. lateralis* (Hig. et Bouck.), *D. japonicus* (Igo et Koike), *D. noduliferus* (Ell. et Grav.), and *D. praenoduliferus* Nigm. et Nem. Redeposited *Lochriea cruciformis, Gnathodus bilineatus bollandensis, Lochriea commutatus*, and *Gnathodus texanus* occur throughout the zone. The assemblage of the *Late D. noduliferus* Subzone was also found on the western and eastern slopes of the South Urals (Kulagina *et al.*, 1992, 1997). It corresponds to the assemblage of the *D. noduliferus* Zone in England (Higgins, 1995; Ramsbottom *et al.*, 1979).

Idiognathoides sinuatus Zone

The *Id. sinuatus* Zone extends throughout beds 6–31 of Exposure 4a and all beds of Exposure 4b. The conodont assemblage is represented here by species inherited from the *D. noduliferus* Zone and combined with *Idiognathoides asiaticus* Nigm. et Bem., *Id. corrugatus* Har. et Holl., *Id. sinuatus* Har. et Holl., and with redeposited *Gnathodus bilineatus bollandensis, Gnathodus bilineatus bilineatus, Lochriea nodosus, Lochriea mononodosus, Lochriea commutatus*, and *Polygnathus inornatus*. In Exposure 4a, the lower zone boundary lies within the beds with ammonoids characterizing the lower part of the *Homoceras coronatum* Zone. It is drawn at probably isochronous levels in many sections of the western and eastern slopes of the South Urals (Kulagina *et al.*, 1992, 1997; *Putevoditel'...*, 1995).

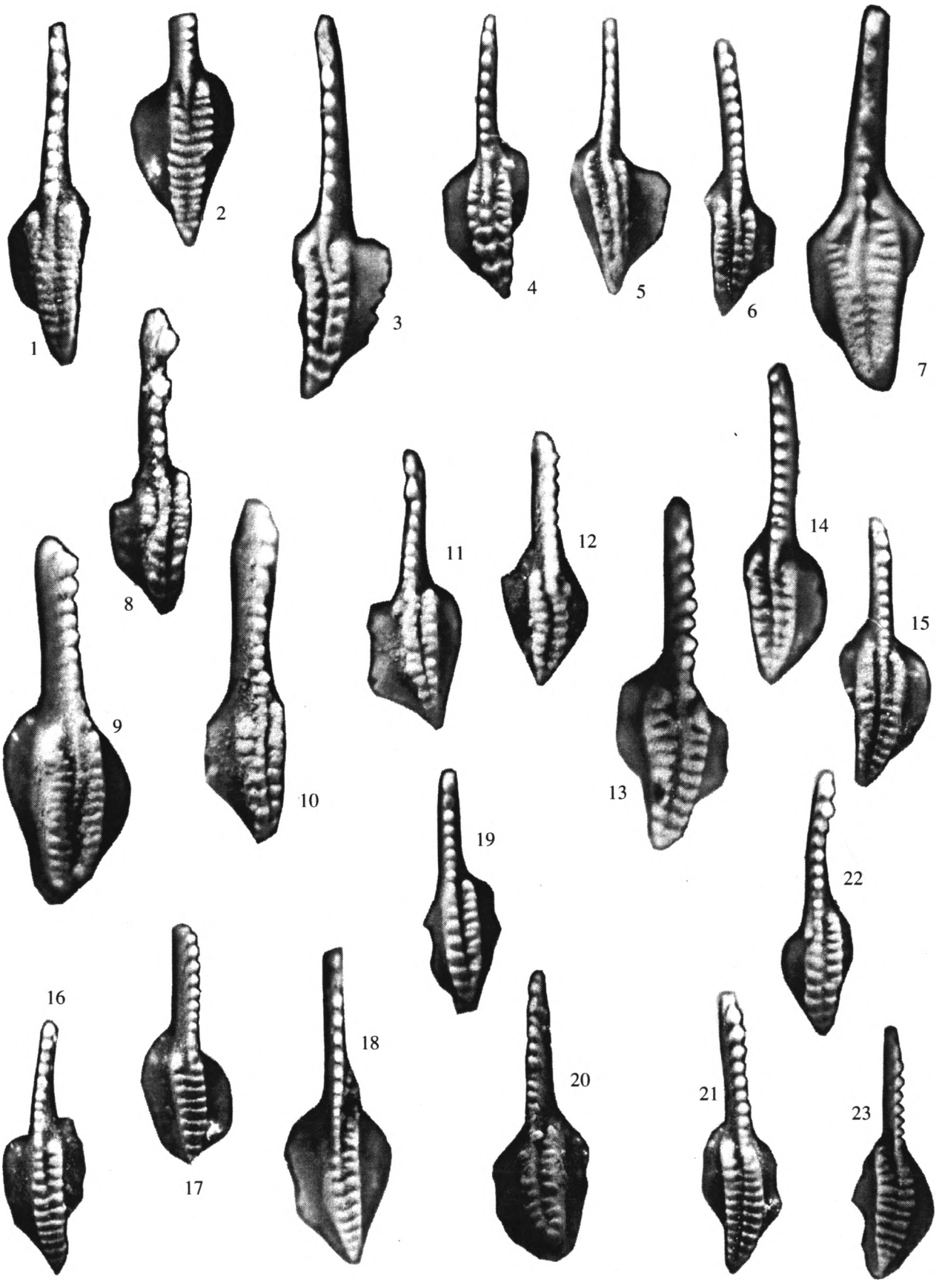
The conodont assemblage is not very comparable to that from the Bashkirian hyperstratotype at the Askyn River, because the last section contains rare conodonts of the shallow-water *Rachystognathus-Declinognathodus* biofacies. The assemblage of the *Early D. noduliferous* Subzone has not yet been found in the lower Bogdanovskii Horizon of the Askyn section, whereas the assemblage of the *Id. sinuatus* Zone appears there at a much higher level than in sections of deep-water deposits (Nemirovskaya and Alekseev, 1995; *Putevoditel'...*, 1995).

FORAMINIFERS

Foraminifers of the Bol'shaya Suren' River were studied by Kireeva, Minyaeva, and Reitlinger (see in Einor *et al.*, 1973a, 1973b). Reitlinger (1980) described

Plate II. Conodonts from the right bank of the Bol'shaya Suren' River, site near the Novyi Bogdanovskii Settlement (×50 for all specimens).

(1–3) *Declinognathodus praenoduliferus* Nigmadganov et Nemirovskaya: (1) Specimen 104/600, Exposure 4A, Sample P-1; (2) Specimen 104/601, Exposure 4a, Sample P-11; (3) Specimen 104/602, Exposure 4b, Sample P-18. (4–6, 14) *Declinognathodus inaequalis* (Higgins): (4) Specimen 104/603, Exposure 4b, Sample P-26; (5) Specimen 104/604, Exposure 4a, Sample P-11; (6) Specimen 104/605, Exposure 4a, Sample P-17; (14) Specimen 104/621, Exposure 4a, Sample P-11. (7, 13, 15) *Declinognathodus lateralis* (Higgins et Bouckaert): (7) Specimen 104/606, Exposure 4b, Sample P-18; (13) Specimen 104/607, Exposure 4a, Sample P-11; (15) Specimen 104/608, Exposure 4a, Sample P-11. (8, 10) *Declinognathodus noduliferus* (Ellison et Graves): (8) Specimen 104/609, Exposure 4a, Sample P-1; (10) Specimen 104/610, Exposure 4a, Sample P-04. (9) *Declinognathodus* aff. *lateralis* (Higgins et Bouckaert), Specimen 104/620, Exposure 4a, Sample P-11. (11, 12, 22) *Declinognathodus japonicus* (Igo et Koike): (11) Specimen 104/611, Exposure 4a, Sample P-04; (12) Specimen 104/612, Exposure 4a, Sample P-1; (22) Specimen 104/622, Exposure 4a, Sample P-7. (16–18) *Idiognathoides asiaticus* Nigmadganov et Nemirovskaya: (16) Specimen 104/613, Exposure 4a, Sample P-7; (17) Specimen 104/614, Exposure 4a, Sample P-03; (18) Specimen 104/615, Exposure 4a, Sample P-11. (19, 20) *Idiognathoides sinuatus* Harris et Hollingsworth: (19) Specimen 104/616, Exposure 4b, Sample P-28; (20) Specimen 104/617; Exposure 4a, Sample P-03. (21, 23) *Idiognathoides corrugatus* Harris et Hollingswort: (21) Specimen 104/618, Exposure 4a, Sample P-7; (23) Specimen 104/619, Exposure 4a, Sample P-03.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the foraminiferal assemblage of the Bogdanovskii Horizon from three sections (Bogdanovka, Yamashly, and Uskalyk). This assemblage contains the characteristic form *Plectostaffella bogdanovkensis*, which was subsequently used as a zonal index species. Kireeva (1978) noted a similarity in the foraminiferal assemblage from the Bogdanovskii Horizon with that of the Krasnaya Polyana Horizon of the Russian platform. According to Sinitsyna *et al.* (1984), who studied the Bashkirian hypostratotype at the Askyn River, the *Pl. bogdanovkensis* Zone corresponds to the lower part of the *Eostaffella pseudostruvei-E. postmosquensis-E. varvariensis* Zone of the general Russian scale. The beds of the Askyn section, which were earlier referred to the Syuran Horizon, were included in the last publication into the Bogdanovskii Horizon. M.B. Postoyalko (see in *Stratigraficheskie...*, 1993) reported that the species *Pl. bogdanovkensis* is associated with the typical "Syuran" assemblage of foraminifers in many sections of the Bashkirian Stage, and, accordingly, the Bogdanovskii Horizon was not accepted as the independent stratigraphic unit. The assemblage of *E. pseudostruvei-E. postmosquensis-E. varvariensis* Zone can be traced up to the top of the Syuran Horizon without being significantly changed. Within this interval, however, the genus *Semistaffella* appeared and evolved. Therefore, it would be more correct to introduce the *Eostaffella pseudostruvei s. l* Zone with three subzones (Kulagina and Sinitsyna, 1997). In the sections studied, foraminifers were found in clotted limestones (biopelsparites), organogenic-clastic limestones (litho-bioclastic packstones and grainstones), and organogenic limestones (bioclastic wackestones and packstones). The clotted limestones (samples P-7, P-9, P-15, P-24, P-25, 172–174, P-41, P-44, P-48, 3/10, and 3/24) yielded an assemblage consisting of very small shells of genera *Mediocris* and *Endostaffella* in association with less frequent *Eostaffella, Plectostaffella, Biseriella*, and *Asteroarchaediscus* forms, whose identification is often difficult. This endosstaffellid biofacies is widespread in both the Serpukhovian and lower Bashkirian deposits. The foraminiferal-detrital limestones (wackestones and bioclastic packstones, samples P-1–P-3, 03, P-10–P-12, P-17, P-18, P-20, P-22, P-24, P-26, 176) of the endothyrid-eostaffellid biofacies bear characteristic species *Tolipammina fortis* Reitl., *Haplophragmina beschevensis* Brazhn., *Pseudoglomospira* spp., *Endothyra bowmani* Phill., *E. phrissa* Zeller, *Semiendothyra surenica* Reitl., *Endotaxis brazhnikovae* (Bog. et Juf.), *Eostaffellina paraprotvae* Raus., *Eostaffella postmosquensis* Kir., *E. pseudostruvei* (Raus. et Bel.) with subspecies, *E. ovoidea* Brazhn. et Pot., *E. pseudoovoidea* Reitl., the group of *E. parastruvei* Raus, *Plectostaffella bogdanovkensis* Reitl., *Pl. reitlingeri* Groves, *Pl. varvariensis* (Brazhn. et Pot.), *Biseriella minima* Reilt., and *Globivalvulina bulloides* (Brady). Less frequent species are *E. (Rectoendothyra) donbassica* Brazhn., *Bradyina cribrostomata* Raus. et Reitl., *Br. concinna* Reitl., archaeodiscids, howchiniids, tetratixiids, and paleotextulariids. The less diverse endothyrid-eostaffellid assemblage (Plate III) is confined to the organogenic-clastic limestones (samples 05B, 04, 03/2, P-13, P-30).

The foraminiferal assemblage described generally corresponds to the *Eostaffella pseudostruvei s.l* Zone consisting of subzones characterized below.

The *Plectostaffella bogdanovkensis* Subzone includes beds 1–9 of Exposure 4a, which are predominantly composed of organogenic-detrital limestones with the rich foraminiferal assemblage of the endothyrid-eostaffellid biofacies. In addition, there occur rare *Plectostaffella evolutica* (Rum.), *Pl.* aff. *varvariensiformis* Brazhn. et Vdov., *Pl. jakhensis* Reitl., *Eostaffella parastruvei* Raus., *E. mirifica* Brazhn., *E. mosquensis* Viss., *Archaediscus krestovnikovi* Raus., *A. vicherensis* (Grozd. et Leb.), *Asteroarchaediscus bashkiricus* Krest. et Theod., *Ast. rugosus* (Raus.), *Neoarchaediscus postrugosus* (Reitl.), *Monotaxinoides tran-*

Plate III. Foraminifers from the right bank of the Bol'shaya Suren' River, site near the Novyi Bogdanovskii Settlement (all specimens, except for those specially indicated, are from Exposure 4).

(1–3) *Eostaffella pseudostruvei* (Rauzer et Belayev), ×85: (1) Specimen 121/847, Sample P-1A; (2) Specimen 121/443, Sample P-18; (3) Specimen 121/848, Sample P-2. (4, 5) *Eostaffella pseudostruvei angusta* Kireeva: (4) ×85, Specimen 121/849, Sample 3b; (5) Specimen 121/446, Sample P-20. (6) *Eostaffella ovoidea* Brazhnikova et Potiyevskaya, ×75, Specimen 121/438, Sample P-18. (7) *Eostaffella pseudoovoidea* Reitlinger, ×85, Specimen 121/850, Sample P-3. (8) *Eostaffella postmosquensis* Kireeva, ×85, Specimen 121/851, Exposure 3, Sample 3/13. (9) *Eostaffella postmosquensis acutiformis* Kireeva, ×85, Specimen 121/852, Sample P-10. (10) *Eostaffella nauvalia* Rumjanzeva, ×85, Specimen 121/853, Sample P-1A. (11) *Millerella paraumbilicata* Manikalova, ×85, Specimen 121/854, Sample P-20. (12) *Plectostaffella evolutica* (Rumjanzeva), ×75, Specimen 121/442, Sample P-26. (13, 14) *Plectostaffella varvariensis* Brazhnikova et Poliyevskaya, ×85, Exposure 3, Sample 3/25; (13) Specimen 121/855; (14) Specimen 121/856. (15, 16) *Plectostaffella* ex gr. *varvariensis* Brazhnikova et Potiyevskaya, ×75, (15) Specimen 121/456, Sample 176; (16) Specimen 121/857, Sample P-20. (17) *Plectostaffella* sp., ×85, Specimen 121/858, Sample 172. (18, 19) *Plectostaffella bogdanovkensis* Reitlinger: (18) ×85, Specimen 121/859, Sample P-1A; (19) ×75, Specimen 121/453, Sample P-20. (20) *Plectostaffella jakhensis* Reitlinger, ×85, Specimen 121/860, Exposure 3, Sample 3/14. (21) *Plectostaffella obtusa* Reitlinger, ×75, Specimen 121/837, Exposure 3, Sample 3/10. (22) *Plectostaffella* aff. *varvariensiformis* Brazhnikova et Vdovenko, ×85, Specimen 121/861, Sample P-2. (23) *Semistaffella?* sp., ×100, Specimen 121/862, Sample P-2. (24, 25, 28, 29) *Semistaffella minuscilaria* Reitlinger, ×100: (24) Specimen 121/863, Sample 03; (25) Specimen 121/864, Sample 172; (28) Specimen 121/865, Sample P-11; (29) Specimen 121/866, Sample P-45. (26, 27) *Plectostaffella reitlingeri* Groves, ×100: (26) Specimen 121/867, Sample P-3B; (27) Specimen 121/868, Sample 03. (30) *Pseudostaffella* aff. *primitiva* Reitlinger, ×100, Specimen 121/869, Sample P-47. (31) *Pseudostaffella antiqua* (Dutkevich), ×85, Specimen 121/870, Exposure 3, Sample 3/31. (32) *Pseudostaffella uralica* Kireeva, ×85, Specimen 121/871, Sample 3/31.

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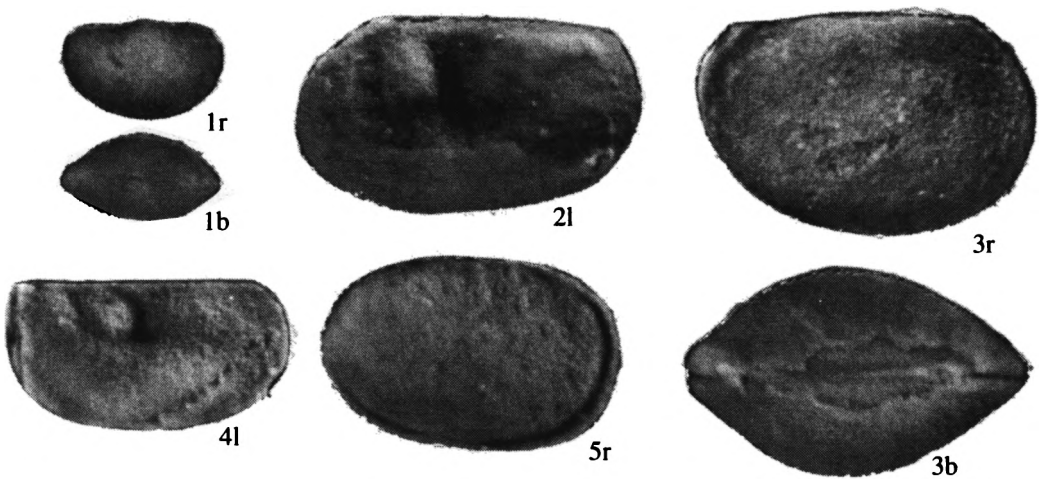

Plate IV. Middle Carboniferous ostracodes from the Mt. Kamennaya site (all specimens are from Exposure 4a located to the southeast of the Novo-Bogdanovskii Settlement, ×50):

(1) *Ardmorea gibberosa* (Knight), Specimen 66-53, Sample 04/1; (2) *Perprimitia digna* N. Kochetova, Specimen 66-210, Sample P-11; (3) *Chamishaella opima* Kotschetkova; (4) *Limnoprimitia arcuata* (Bean), Specimen 66-26, Sample 04/1; (5) *Javatis kisilensis* (Kochetkova), Specimen 66-1, Sample 03. Symbols: (r) right valve view, (l) left valve view, (b) dorsal view.

sitorius Brazhn. et Pot., *Eolasiodiscus donbassicus* Reitl., single *Howchinia gibba* (Moell.), and *Mesolasiodiscus?* sp. Paleotextulariids and tetrataxiids are also present. This zone is comparable in stratigraphic range with beds 5–7 of the Bashkirian stratotype at the Askyn River (Sinitsyna and Sinitsyn, 1987).

The *Semistaffella minuscilaria* Subzone corresponds to beds 10–37 of Exposure 4a, beds 10–47 of Exposure 4b, and, probably, to beds 1–6 of Exposure 3. In Exposure 4a, the dominant species are those inherited from the underlying deposits. *Millerella paraumbilicata* Man. and *Globivalvulina kamensis* Reitl. appear there for the first time. Some interbeds (samples 10, 18) contain *Parastaffella struvei* (Moell.), *Par. utkaensa* Post., *Pseudoendothyra illustria* (Viss.), and *Ps. schlykovae* (Durk.). Beds 29–47 of Exposure 4b and beds 1–11 of Exposure 3 are not rich in foraminifers. Rare interbeds of fine-grained detrital limestones present in this interval yielded species characteristic of the *Eostaffella pseudostruvei* Zone. The subzone under consideration is correlative with the interval of the Askyn section from the upper part of Bed 7 to Bed 11 (Kulagina and Sinitsyna, 1997). According to occurrence of *Semistaffella* aff. *primitiva* Reitl. (Sample 47), beds 48–49 of Exposure 4b presumably correspond to the *Semistaffella variabilis* Zone. The beds are predominantly composed of pelitomorphic limestones lacking foraminifers. Rare interbeds at this level contain *Eostaffella postmosquensis, Millerella paraumbilicata, Plestostaffella* sp., and *Semistaffella* sp. Beds 7–11 of Exposure 3 with scarce foraminifers can be attributed to the same subzone. In Exposure 3, the considerable unexposed interval is succeeded by the limestone outcrops, where we identified *Pseudostaffella antiqua* (Dutk.), *Ps. uralica* Raus., *Ps. paracompressa* Raus., and other species characteristic of the Akavas and overlying horizons of the Bashkirian Stage. Taking into consideration the thickness of the Syuran Horizon in this structural-facial zone, we believe that its upper boundary is between beds 12 and 13.

The foraminiferal zonation of the Syuran Horizon is well correlative with that established in the central Tien Shan (Rumyantseva, 1989). The described assemblages are very similar to those known in the Donets basin and North America (Table 2).

OSTRACODS

The stratotype of the Bogdanovskii and Syuran horizons near the Novyi Bogdanovskii Settlement is characterized by the ostracode assemblage of the *Ardmorea gibberosa-Limnoprimitia arcuata* Zone recognized by Kochetkova (1983) in many sections of the South Urals. The assemblage consists of more than 50 species from 31 genera and 18 families of paleocopids, cladocopids, platicopids, metacopids, and podocopids (Plate IV). The lowermost beds of the Syuran horizon (Exposure 4a, samples 05-B, 05-A, P-1/B) yielded the paraparchytid-bairdiid assemblage of scarce shells of *Polycope perminuta* (Kellett), *Shishaella claytonensis* (Knight), *Shis. circinata* N. Kotch., *Shivaella evidens* Kotsch., *Chamishaella uniformis* Kotsch., *Microcoeloenella orbiculata* Kotsch., *Healdia procerula* Kotsch., *Bairdia alula* Kotsch., *B. laklyensis* Kotsch., *B. chudolasensis* Kotsch., and *Acratia* cf. *grandis* Kotsch. The overlying beds (Exposure 4a, sample 0-4/1 and others from higher levels; Exposure 4b, samples P-18, P-25, P-26, P-39, P-41) contain the most representative assemblage of the zone, which includes, along with the forms listed above, *Javatius eximius* Kotsch.,

Table 1. Relationships between ammonoid, foraminifer, conodont, and ostracode zonations in the Syuran Horizon of the South Urals

Series	Stage	Horizon	Ubhorizon	Ammonoid genus-zone	Ammonoid zone of West Europe	Zones, subzones, beds: foraminifers		conodonts		ostracodes	Sections: Novo-Bogdanovskii Exposure 3	Novo-Bogdanovskii Exposure 4	Muradymovo	Askyn
Middle Carboniferous	Bashkirian	Akavas		*Bilinguites-Cancelloceras*	G1	*Pseudostaffella antiqua*		*Neognathodus askynensis*		*Coryellina inflata-Kirkbyella aperta*				
					R2									
		Syuran	?	*Reticuloceras-Bashkortoceras*	R1	*E. pseudostruvei*	*Semistaffella variabilis*	*Idiognathoides sinuatus*		*Ardmorea gibberosa-Limnoprimitia arcuata*				
					H2		*Semistaffella minuscularia*							
			Bogdanovskii	*Homoceras-Hudsonoceras*	H1		*Plectostaffella bogdanovkensis*	*D. nodulif.*	*Late*					
						E. explicata M. subplana	*Plectostaffella varvariensis*		*Early*	*Fellerites gratus* Beds				
Lower Carboniferous	Serpukhovian	Staroutkinskii		*Fayettevillea-Delepinoceras*	E2		*Monotaxinoides transitorius*	*Gnathodus bilineatus bollandensis*		*Carbonita ? subquadrata-Aurigerites solitarius Anachuacia rara*				

Coeloenellina ultima Kotsch., *C. serotina* Kotsch., *Glyptopleura* cf. *subvarians* Kotsch., *Idiomorphina subsimplex* N. Kotch., *Polycope? rugosa* Kotsch., *Discoidella perpicua* Kotsch., *Shivaella asselica* Jag., *Bairdia rustica* Kotsch., *B. bogdanovkaensis* N. Kotch., *B. gibbus* Kotsch., *Bairdiacypris subconspicuus* Kotsch., *B. indiges* Kotsch., *B. obtusus* Kotsch., *Acratia demissa* Kotsch., *Roundyella subaculeata* N. Kotch., *Macrocypris lenticularis* Cooper, and *M.* cf. *menardensis* Harlton, as well as *Ectodemites planus* Cooper, *Kirkbuina tenella* N. Kotch., *Bolbozoella inflata* Grundel, *Rectonaria accepta* N. Kotch., *Triplacera imperspicua* N. Kotch., and *Basslerella longa* N. Kotch. These species are typical of the upper Serpukhovian deposits. The presence of the genera *Kirkbyella, Coryellina, Cavellina,* and *Paraberounella* is remarkable. Also indicative is the coexistence of index-species *Limnoprimitia arcuata* (Bean) and *Ardmorea gibberosa* (Knight) with single *Javatius kisilensis* (Kotsch.), *Perprimitia digna* N. Kotch., and *Chlamishaella opima* Kotsch.

The upper part of Exposure 4b (Sample 49) and beds 5–11 of Exposure 3 (samples 3/10, 3/13, 3/24, 3/25) yielded an impoverished assemblage of the *Ard. gibberosa-L. arcuata* Zone, which includes rare *Perprimitia* sp., *Glyptopleura* sp., *Microcoeloenella orbiculata* Kotsch., and *Dorsoobliquella ovalis* Kotsch. Exposure 3 also yielded *Coryellina* sp., *Kirkbya* sp., *Limnoprimitia* cf. *arcuata* (Bean), *Idiomorphina* sp., and *Glyptolichwinella postuma* Kotsch. In association with poorly preserved and incomplete shells of paraparchytids and bairdiids, few of which are identifiable at the species level.

On the basis of ostracodes, the Syuran Horizon is correlative with coeval deposits of the central Tien Shan, where two assemblages characterize the *Perprimitia digna* and *Javatius kisilensis-Chlamishaella opima* beds. The described assemblages are similar in species composition to that of the *Ard. gibberosa-L. arcuata* Zone of the Urals, but they lack the index-species of the latter.

DISCUSSION

The detailed investigations in the type area of the Syuran and Bogdanovskii horizons showed that both can be recognized in the same limestone succession exposed at the left bank of the Bol'shaya Suren' River. The Bogdanovskii Horizon corresponds to the lower part of Exposure 4 (according to the terminology of Einor *et al.*, 1973a, 1973b), and the Syuran Horizons to the upper part of Exposure 4 and to Exposure 3 (Table 1). The Bogdanovskii Horizon, as established by Einor *et al.* (1973a, 1973b), actually represents the lower part of the Syuran Horizon and is equal in the stratigraphic range to the ammonoid *Homoceras* Zone

Table 2. Correlated foraminiferal zonations in different regions

<table>
<tr><th>General foraminiferal zonation in Russia (Reshenie..., 1990)</th><th colspan="2">The Urals (Stratigraphicheskie..., 1993), western and eastern subregions</th><th colspan="2">South Urals (Kulagina and Sinitsyna, 1997)</th><th>Central Tien Shan (Rumyantseva, 1989)</th><th>Donbas (Vdovenko et al., 1989)</th><th>North America, Arrow Canyon (Brenckle et al., 1997; Ross, 1997)</th></tr>
<tr><td>Ps. antiqua</td><td colspan="2">Ps. antiqua</td><td colspan="2">Ps. antiqua</td><td>Ps. antiqua</td><td>Ps. antiqua</td><td>Pseudostaffella sp.</td></tr>
<tr><td rowspan="2">Pl. varvariensis
E. pseudostruvei
E. postmosquensis</td><td colspan="2" rowspan="2">E. postmosquensis
Pl. jakhensis</td><td rowspan="3">E. pseudostruvei</td><td>S. variabilis</td><td>S. variabilis</td><td rowspan="2">E. postmosquensis
E. pseudostruvei
Pl. varvariensis</td><td rowspan="2">Millerella marblensis</td></tr>
<tr><td>S. minuscillaria</td><td>Pl. seslavica</td></tr>
<tr><td>Pl. bogdanovkensis</td><td rowspan="2">Parastaffella utkaensa–Eoplect. acuminata</td><td rowspan="2">Br. cribrostomata Pl. varvariensis</td><td>Pl. bogdanovkensis</td><td>Pl. bogdanovkensis</td><td>Pl. berestovensis-E. angusta</td><td>Mill. pressa</td></tr>
<tr><td rowspan="2">Eosigmoilina explicata–Monotaxinoides subplana</td><td colspan="2">Pl. varvariensis*</td><td>Pl. posochovae*</td><td rowspan="2">Loeblichia* minima–M. transitorius
Eos. explicata</td><td>Breckleina rugosa*</td></tr>
<tr><td colspan="2">M. transitorius</td><td>Eos. explicata</td><td>Eosigmoilina robertsoni</td></tr>
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* First occurrence level of *Declinognathodus noduliferus*.

(the *Homoceras-Hudsonoceras* Genus-zone of the Carboniferous stratigraphic scale of Russia). It was long deemed that the *Reticuloceras-Bashkortoceras* Genus-zone spans the greater part of Exposure 4, while the *Homoceras-Hudsonoceras* Genus-zone is restricted to the rest small portion of the section (beds 16–34). Beds 1–15 were distinguished in the recent prospecting trench (Kulagina *et al.*, 1992). However, recent ammonoid findings suggest that the Bogdanovskii Horizon should be extended upward throughout the 34-m-thick overlying beds. In the stratotype of the Bogdanovskii Horizon, its lower boundary drawn at the base of the ammonoid *Homoceras* Genus-zone is not observable, and the lowermost beds corresponding to the ammonoid *Isohomoceras subglobosum* Zone and to the conodont *Early D. noduliferus* Subzone are missing. This interval was recognized in some sections of western and eastern slopes of the South Urals. It is completely represented in the Muradymovo Village section correlative with the stratotype and located in the same structural-facial zone (Kulagina *et al.*, 1992; *Putevoditel'...*, 1995).

In the Muradymovo section, the lower boundary of the Syuran Horizon is placed at the base of the *Early D. noduliferus* Subzone, where conodonts *D. noduliferus* (Ell. et Grav.) and *D. inaequalis* (Hig.) appear. The assemblage of the subzone is dominated by species that first appeared in Visean and Serpukhovian times, such as *Gnathodus bilineatus bilineatus* (Roundy), *Gn. bilineatus bollandensis* Hig. et Bouck., *Gn. postbilineatus* Nigm. et Nem., *Lochriea commutata* (Br. et Mehl), *L. costata* (Paz. et Nem), *L. monocostata* (Paz. et Nem.), *L. mononodosa* (Rh., Aust. et Dr.), *L. nodosa* (Bisch.), and *L. ziegleri* Nem., Per. et Meisch. Less frequent are *Cavusgnathus uniformis* Youngq. et Mill. and *Mestognathus bipluti* Hig. The subzone can be correlated with lower beds of the Voznesensk Horizon of the Donets basin (Limestones D58–D59), which contains *D. noduliferus* and *Gn. bilineatus bollandensis* (*Verkhneserpukhovskii...*, 1983; Nemirovskaya and Nigmadzhanov, 1994). In the southern Tien Shan, the Suffa Formation of the Gissar Ridge (the Aksu 1 and Aksu 4 sections) includes an interval, which yields both the Serpukhovian and Bashkirian conodonts (Nigmadganov and Nemirovskaya, 1992; Nemirovskaya and Nigmadganov, 1994). In contrast to the Urals, *D. praenoduliferus* appears in Tien Shan at a lower level than other representatives of the genus *Declinognathodus* being almost concurrent with *Idiognathoides asiaticus* and *I. sulcatus*. In the stratotype of the mid-Carboniferous boundary (the Arrow Canyon section), the lower *D. noduliferus* Subzone is confined to the shallower marine biofacies characterized by representatives of the genera *Declinognathodus, Adetognathus, Rhachistognathus*, and *Gnathodus* (the group of *girtyi* and *deflectus* after Brenckle *et al.*, 1997). In the stratigraphic schemes of the Urals (*Stratigraficheskie...*, 1993), the lower boundary of the Bashkirian Stage is drawn at the base of the foraminiferal *E. postmosquensis-Pl. jakhensis* Zone corresponding to the *P. bogdanovkensis* and *E. pseudostruvei-E. postmosquensis-Pl. varvariensis* zones of the general stratigraphic scale of Russia. The *Pl. bogdanovkensis* Zone is correlative with the ammonoid *Homoceras* Zone. However, the thorough investigations of sections in the South Urals revealed that *D. noduliferus* appeared for the first time (Table 2) in the beds bearing Serpukhovian foraminifers (Kulagina *et al.*, 1992). The same distribution patterns are recorded in the Donets basin (*Verkhneserpukhovskii...*, 1983; Vdovenko *et al.*, 1989). In the Muradymovo section with abundant Late Serpukhovian species, the earliest *D. noduliferus* is found in association with rare *Plectostaffella varvariensis* Brazhn. and *Planoendothyra aljutovica* Reitl. We recognized the last level as the *Pl. varvariensis* Subzone correlative with the upper part of the *Eosigmoilina*

explicata-Monotaxinoides subplanus Zone of the general scale (Table 1). This zone may correspond to the *Plectostaffella posochovae* Zone (Rumyantseva, 1989; Kulagina *et al.*, 1992) underlying the *Pl. bogdanovkensis* Zone of central Tien Shan.

In the Muradymovo section, the earliest conodonts *D. noduliferus* (beds 9a and 9b) are associated with a transitional ostracode assemblage characteristic, in our nomenclature, to the *Fellerites gratus* Beds. The assemblage includes the rare early Bashkirian forms, such as *Bairdia chudolasensis* Kotsch., *Bairdiacypris indiges* Kotsch., *Basslerella simonovae* Kotsch., and *Bass. subcrassa* Kotsch., all associated with predominant Serpukhovian species *Kirkbyina tenella* N. Kotch., *Healdia uralica* N. Kotch, *H. ikensis* N. Kotch., *Bolbozoella nodosa* Robinson, *Bolb. inflata* Grundel, and *Bairdiocypris subalia* N. Kotch., and also with a peculiar facies-restricted association of kirkbyaceans consisting of representatives of genera *Kirkbya, Amphissamples, Ectodemites, Amphizona*, and *Kellettina*. Rare specimens of *Fellerites gratus* N. Kotch. et Vak. are characteristic of this level as well. The overlying bed yielded scarce ostracode shells characteristic of the *Admorea gibberosa-Limnoprimitia arcuata* Zone, its index-species included (Kulagina *et al.*, 1992). Thus, the Syuran Horizon corresponds to two ammonoid *Homoceras-Hudsonoceras* and *Reticuloceras-Bashkortoceras* genus-zones, to two conodont zones of *Declinognathodus noduliferus* (with the *Early D. noduliferus* and *Late D. noduliferus* subzones) and *Idiognathoides sinuatus*, to the foraminiferal *Plectostaffella varvariensis* Subzone and *Eostaffella pseudostruvei* Zone (with three subzones), and to the *Fellerites gratus* Beds and *Ardmorea gibberosa-Limnoprimitia arcuata* Zone of ostracodes (Table 1). The important change in the conodont assemblages (the appearance of *Idiognathoides sinuatus*) almost coincides in time with the appearance of *Semistaffella minuscilaria* within Zone H_1. The Syuran Horizon comprises the greatest number of ammonoid genus-zones as compared to the Serpukhovian and Bashkirian horizons corresponding either to a one genus-zone or to a part of such subdivision. The Bogdanovskii Horizon spanning the range of the *Homoceras-Hudsonoceras* Genus-zone and representing a part of the Syuran Horizon can be incorporated as a subhorizon into the stratigraphic scheme of the Urals. According to our data, the Bogdanovskii Horizon is about 44 m thick in the stratotype. The overlying deposits of the *Reticuloceras-Bashkortoceras* Genus-zone should be recognized as a new subhorizon, whose lower boundary is coincident with the lower boundary of the ammonoid *Surenites krestovnikovi* Zone is probably located close to that of the foraminiferal *Semistaffella variabilis* Subzone.

The problem of recognizing a new horizon in the considered part of the section was discussed in earlier publications (Nikolaev, 1989; Einor, 1992). The problem of positioning the upper boundary of the Syuran Horizon on the basis of ammonoid zonation remains open at present, unless the ammonoid succession will be adequately studied in the basal beds of the overlying Akavas Horizon.

CONCLUSIONS

(1) The stratotype of the Syuran Horizon does not exhibit its lower boundary coincident with the base of the Middle Carboniferous. The stratotype lacks the lowermost beds of the horizon, which correspond to the ammonoid *Isohomoceras subglobosum* Zone, to the conodont *Early D. noduliferus* Subzone, to the foraminiferal *Pl. varvariensis* Subzone, and to the ostracode *Fellerites gratus* Beds. This interval is completely represented in the section near the Muradymovo Village, which is suggested as the hypostratotype of the Syuran Horizon.

(2) The Syuran Horizon corresponds to the following biostratigraphic units: the ammonoid *Homoceras-Hudsonoceras* and *Reticuloceras-Bashkortoceras* genus-zones; the conodont *D. noduliferus* Zone (with two, Late and Early Subzones) and *Id. sinuatus* Zone; the foraminiferal *Pl. varvariensis* Subzone (the upper part of the *Monotaxinoides subplanus-Eosignoilina explicata* Zone), the *E. pseudostruvei* Zone with three subzones (*Pl. bogdanovkensis, S. minuscilaria*, and *S. variabilis*); the *Fellerites gratus* Beds and *Ardmorea gibberosa-Limnoprimitia arcuata* Zone of ostracode zonation.

(3) The Bogdanovskii Horizon actually represents the lower part of the Syuran Horizon and should be included into the Unified stratigraphic scheme of the Urals as the subhorizon equal in scope to the ammonoid *Homoceras-Hudsonoceras* Genus-Zone. In the stratotype of the Bogdanovskii Subhorizon, the *Homoceras-Hudsonoceras* Genus-Zone is characterized by ammonoids of the *Homoceras coronatum* Zone; whereas the *Hudsonoceras proteum* Zone is conventionally recognized by the presence of the fine-sculptured *Homoceras* forms.

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The Eifelian Vegetation and Reconstructed Landscapes of Central Kazakhstan

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Abstract—The taxonomic composition and structure of Eifelian plant communities of central Kazakhstan and the orogeny-influenced distribution of taxa are discussed. Circular diagrams are plotted for individual floral taxa from sedimentary formations of strictly continental genesis and from those including marine sediments. The data obtained, as well as plant life forms and a relief differentiation, are taken into account in the reconstruction of Eifelian landscapes of the Devonian volcanic belt and some adjacent Hercynides areas. The higher and dissected relief of the volcanic belt was characterized by more abundant and diverse vegetation than the deltaic marshes or flooded terraces in the Hercynides areas. In both cases, there was no continuous vegetation cover and the plants grew in isolated areas, which were either damp or located near water basins.

Key words: *central Kazakhstan, Eifelian, continental deposits, flora, landscape reconstruction*

INTRODUCTION

The Eifelian floral assemblage of central Kazakhstan consisting of four species (*Protolepidodendron scharianum* Krejci, *Blasaria sibirica* (Krysht.) Zal., *Calamophyton* sp., and *Dicranophyton niayssiense* Zal.) was first identified by Senkevich (1957, 1959). *Blasaria sibirica* (Krysht.) Zal. abundant in marine facies of the Northern Balkhash region was considered to be the predominant species. Other species were distinguished in the northern part of central Kazakhstan, where the enclosed deposits were dated according to their stratigraphic position. Later, the Eifelian floral assemblages from predominantly marine deposits were discussed in the following publications: Senkevich (1982), Kaplun and Senkevich (1971), Senkevich and Kaplun (1982), and Dubatolov *et al.* (1977). Among other geological problems of central Kazakhstan, the recognition of the Eifelian Stage, especially in continental facies, was the most challenging. This problem was put forward as early as the 1960s by Bogdanov *et al.* (1960). At that time, investigations were being carried out in the southern Karaganda basin, where the Eifelian deposits were conventionally recognized by the stratigraphic position. It was supposed that the Eifelian time was marked by the accumulation of the lower sequence of basaltic andesites, because they lie in a continuous volcanogenic section between horizons with the Early Devonian and Givetian floras. It was also believed that the accumulation event occurred in the eastern part of the Karaganda basin, while in the western part lacking the Eifelian deposits, there was a break in sedimentation in response to the tectonic elevation. Bogdanov persistently pointed to the significance of discovering the continental fossiliferous deposits of the Eifelian age. Later it became clear that the problem of the Eifelian Stage recognition is important not only for southern margins of the Karaganda basin, but also for many other Paleozoic structures of central Kazakhstan. In addition to different-scale geological mapping, Bogdanov organized special stratigraphic and paleobotanical works. The latter progressed most intensively, because the continental deposits are often characterized only by plant remains. The studies under Bogdanov's supervision were made by researchers of the central Kazakhstan Expedition of Moscow State University: O.E. Belyaev, E.A. Bilenko, B.Ya. Zhuravlev, V.N. Zavrazhnov, Yu.F. Kabanov, A.L. Kotlyar, O.A. Mazarovich, S.P. Malinovskaya, I.A. Poslavskaya, A.V. Ryazantsev, A.P. Uryvaeva, G.T. Ushatinskaya, T.O. Fedorov, N.P. Chetverikova, N.L. Chechik, L.A. Shchegoleva, M.N. Shcherbakova, and A.L. Yurina. Some other geologists (M.K. Bakhteev of Moscow Geological-Prospecting Institute, Moscow; Yu.A. Vasyukov of Centrkazgeologiya, Karaganda; and V.M. Shuzhanov of Institute of Geological Sciences of KazSSR, Alma-Ata) also contributed much to the problem. Yurina performed the earliest investigations of the Devonian flora in continental deposits of central Kazakhstan.

STRATIGRAPHY

In the Devonian volcanic belt (DVB) of Kazakhstan and in some northern superimposed troughs of the Caledonides, there are continuous volcanogenic and volcanogenic-sedimentary sections, the lower part of which contains the Early Devonian floral assemblage, whereas the upper one yields the Givetian plant

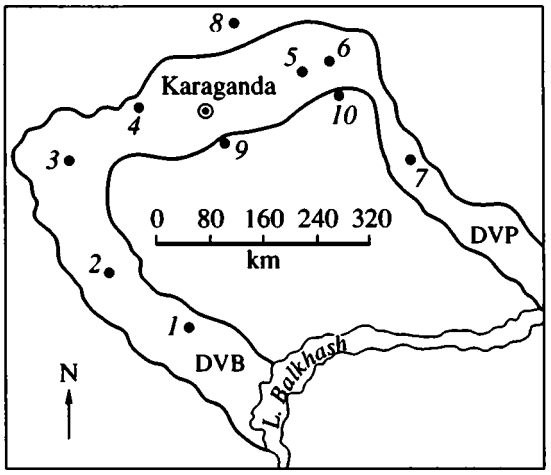

Fig. 1. Floral localities in the Eifelian formations of central Kazakhstan: (*1*) the Degrez Formation; (*2*) the Uguztau Formation; (*3*) the Taldysai Formation; (*4–6*) the Zhilandybulak Formation; (*7*) the Irgaily Formation; (*8*) the Kurtozek Formation; (*9*) the Karabastau Formation; (*10*) the Krykoshak Formation; (DVB) Devonian volcanic belt.

remains. This suggested that the Eifelian deposits must be present here as well, and special investigations were needed to discover floral remains of this age. The early collections included plant remains that were either poorly preserved, or displayed a wide stratigraphic range (the entire Middle Devonian). The Second Conference on Precambrian and Paleozoic Stratigraphy of Kazakhstan, in the fall of 1971, accepted the Eifelian Stage as corresponding to the Irgaily Formation in the southeastern segment of the DVB, to lower parts of the Zhilandybulak and Salkintau formations in the latitudinal segment of the DVB, and to the upper part of the Komadyr (Karabastau in later nomenclature) Formation of the northern Nura synclinorium (Chetverikova *et al.*, 1971) according to constraints of floral identifications, stratigraphic position, and lithological correlation.

In the last decades, after the premature demise of Bogdanov, abundant floral remains were collected during detailed geological mapping and special investigations. This enabled identification of the floral assemblage from the Eifelian continental deposits of central Kazakhstan. By the time of the Third Conference on Kazakhstan Stratigraphy (1986), the Eifelian continental flora-bearing deposits have been discovered and correlated in the DVB and adjacent parts of the Caledonides and Variscides (Mazarovich *et al.*, 1977, 1985; Malinovskaya and Yurina, 1975, 1978, 1983; Yurina *et al.*, 1982).

After the conference and investigations carried out in the DVB and its northern peripheral parts, the diversity and abundance of the Eifelian floral assemblage was clarified (Zhuravlev *et al.*, 1989). Correlation between the continental formations of the DVB latitudinal segment and marine sections in the adjacent northern Karasor synclinorium was substantially improved. The coastal-marine deposits of the latter were thoroughly studied near the Lake Saumalkol and the Krykoshak Mountain, where the Krykoshak Formation similar in lithology to the Zhilandybulak and Irgaily formations was recognized (Belyaev *et al.*, 1996). Biostratigraphic subdivision of these deposits and correlation of the formations on the basis of the Eifelian floral assemblages were discussed in detail by Yurina (Yurina, 1988; Malinovskaya and Yurina, 1975, 1978, 1983). Senkevich and Yurina (see in Bakhteev *et al.*, 1984) elaborated the Eifelian floral zonation of Kazakhstan, but the type of the Eifelian vegetation of central Kazakhstan has not been reconstructed. This article presents the first attempt in this direction. We shall consider floras of the following units exposed in different marginal segments of the Devonian volcanic belt (Fig. 1): Degrez, Uguztau, and Taldysai (the southwestern segment), Zhilandybulak (the latitudinal segment), and Irgaily (the northeastern segment) formations. Flora of the Kurtozek Formation of the Kazakhstan-Tien Shan Caledonides is also considered. The formations listed above are continental in origin. In addition, floras of the Krykoshak Formation and lower Karabastau Subformation of the Nura synclinorium (the Dzhungar-Balkhash Hercynian area of predominantly marine sedimentation) are subjected to analysis. The recognition of six continental formations with the Eifelian flora shows that the problem put forward by Bogdanov has been solved by his younger colleagues.

TAXONOMIC ANALYSIS

To characterize a floral assemblage of each Eifelian formation in general, we should determine quantitative proportions between its species. It should be noted that the relative abundances could be different in the assemblages of similar species composition. To some extent, this may reflect distinctions between some ecological groupings. Yurina performed the quantitative analysis for every heterochronous locality from the very beginning (1953) of her study of the Devonian flora of Kazakhstan. Calculations were somewhat difficult in the case of floral remains collected by other geologists. In this case, the relationships between the collected species were dependent on some attendant circumstances, such as time of collection, the ability of a geologist to recognize floral imprints in sections, and some others. This particularly concerns flora of the Degrez Formation, which was inaccessible for our personal sampling; thus, the corresponding data are of limited use.

The Eifelian plants of the central Kazakhstan areas under consideration include representatives of algae, rhyniophytes, lycopsides, cladoxyleans, progymnosperms, and plants of unclear systematic position (*incertae sedis*). When plotting circular diagrams, these plant categories were classified into ten groups, most of which are significant for the Eifelian biostratigraphy of central Kazakhstan (Fig. 2). The diagrams show the

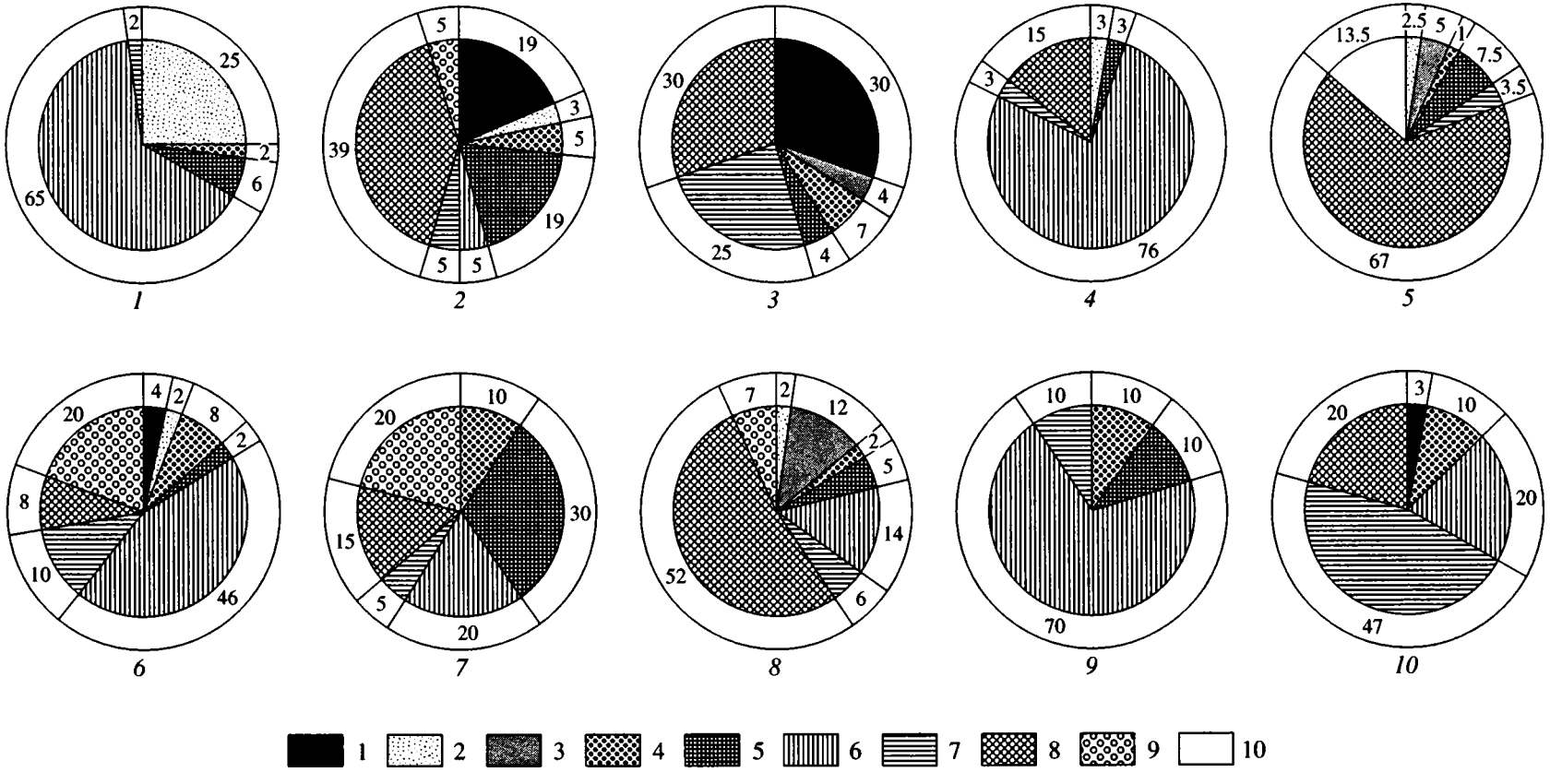

Fig. 2. Circular diagrams showing percentages of floral taxa in the Eifelian formations of central Kazakhstan: (*1*) algae; (*2*) *Parapsilophyton balkhashense*; (*3*) *Kaplunella lissa*; (*4*) other rhyniophytes; (*5*) *Artschaliphyton unicum*; (*6*) *Protolepidodendron scharianum*; (*7*) *Prelepidodendropsis kornilovae*; (*8*) other lycopsides; (*9*) cladoxyleans and progymnosperms; (*10*) plants of uncertain systematic position. Figures in segments denote percentages of taxa relative to the total abundance of species in the assemblage; numbers under circular diagrams correspond to those of the Eifelian formations shown in Fig. 1.

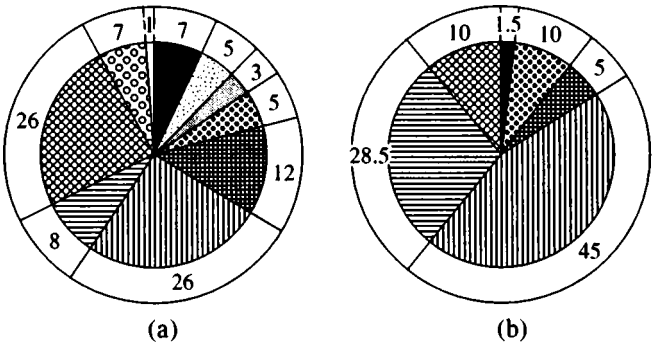

Fig. 3. Generalized circular diagrams for the Eifelian floras from (a) continental and (b) marine deposits of central Kazakhstan (symbols as in Fig. 2).

percentages of a species of each group in the assemblage. We united all algae into one group and distinguished three groups of rhyniophytes: two of *Kaplunella lissa* Senk. and *Parapsilophyton balkhashense* Senk. (these are zonal species of the Eifelian deposits of Kazakhstan, see Bakhteev *et al.*, 1984), and the group consisting of genera *Taeniocrada, Pectinophyton, Sawdonia, Psilophytites*, and *Hostinella*. Lycopsides represent four groups, three of *Artschaliphyton unicum* Senk. (zonal species), *Protolepidodendron scharianum* Krejci (widespread throughout the Middle Devonian), and *Prelepidodendropsis kornilovae* Senk. (characteristic of the Middle Devonian, but known as restricted in distribution to Kazakhstan), and one including genera *Lidasimophyton, Sugambrophyton, Blasaria, Barrandeina*, and *Lycopodites*. Cladoxyleans (genera *Calamophyton, Pseudosporochnus*) and progymnosperms (genus *Rellimia*) constitute a single group. The genus *Arganatiella* of an uncertain systematic position (*incertae sedis*) and characteristic especially of the lower Eifelian deposits is also distinguished as a single group.

Despite a certain similarity, the assemblages of continental formations (CA) are quite different from those of continental formations including marine beds (MA). The principal difference is that the former comprise all the mentioned plant categories showing high diversity, whereas the latter contain only three categories: algae, rhyniophytes, and lycopsides. Comparing individual and generalized circular diagrams characterizing two genetic types of floras (Fig. 3), one may conclude the following. Proportions between the plant groups CA and MA are the same in general; i.e., lycopsides are dominant in both cases (66–80% with average values of 72% in the CA, and 87–90% with an average value of 88% in the MA). This demonstrates the general similarity between floras from genetically different deposits; however, percentages of particular lycopside groups are variable. For instance, *Protolepidodendron scharianum* dominates in the MA (45% in average and up to 70% in the lower Karabastau Formation) and is less abundant in the CA (only 26% on average, despite a local abundance of up to 76% in the Zhilandybulak Formation). The zonal species *Artschaliphyton unicum* represents 2–30% in every CA (12% in average) but not more than 5% in the MA. Other lycopsides correspond to five genera (26% in average) in the CA and to two genera *Blasaria* and *Barrandeina* only (10% in average) in the MA.

Rhyniophytes are scarce in both assemblages and have approximately equal percentages: 10% in the MA and 13% in the CA. They are more diverse in the latter, containing two zonal species with an average abundance level of 3–5% each. An exception is the assemblage of the Degrez Formation, 25% of which is constituted by the zonal species *Parapsilophyton balkhashense*. However, as mentioned above, the data on this assemblage should be used with reservation. Another group of rhyniophytes (five genera) occurs in all DVB formations and in the Kurtozek Formation of the Caledonides, but these plants represent only 1–10% of assemblages (5% in average). The rhyniophyte assemblages from the Krykoshak and lower Karabastau formations lack the zonal species and include only two genera, *Psilophyton* and *Hostinella*, whose abundance is up to 10%.

Algae and related plants are typical of both types of assemblages in subordinate amounts: 1.5% in the marine formations and about 7% in the continental ones.

Cladoxyleans and progymnosperms were conventionally considered as a single group for the convenience of cyclogram plotting. Cladoxyleans are peculiar plants commonly referred to Pteropsida or to Sphenopsida. In our opinion (Yurina, 1988), these plants form an independent group of the Pteridophyta Division and most likely represent a blind evolutionary branch. The class of progymnosperms is unanimously assigned to pteridophytes and supposed to be ancestral for many gymnosperms. Cladoxyleans and progymnosperms are leading groups in the younger (Givetian, Famennian, and Early Carboniferous) floras. In the Eifelian deposits of central Kazakhstan, they show their first occurrence. We found them in three DVB formations (Uguztau, Zhilandybulak, and Irgaily) and in the Kurtozek Formation of the Caledonides, i.e., in deposits of the pure continental genesis. The noteworthy fact is that all DVB formations listed above contain only cladoxyleans (5–20%), while cladoxyleans of the Kurtozek Formation (6%) are associated with early progymnosperms (1%).

From the taxonomic analysis, we arrived at the following general conclusion. The Eifelian assemblages of both marine and continental deposits are rich in lycopsides. The CA includes, along with older forms of rhyniophytes flourishing in the Early Devonian, a small amount (up to 7%) of younger forms of cladoxyleans and progymnosperms, whereas the MA is completely lacking in them. There are seven plant species of algae,

Taxonomic composition of the Eifelian plants from central Kazakhstan

Species	Continental deposits						Marine deposits	
	Kurtozek Formation	Degrez Formation	Uguztau Formation	Lower Taldysai Subformation	Zhilandybulak Formation	Irgaily Formation	Krykoshak Formation	Karabastau Formation
Algae								
Krupchenkella katbasensis Senk.							+	
Orestovia sp.			+		+			
Thamnocladus sp.				+				
Rhyniophytes								
Taeniocrada decheniana (Goepp.)Kr. et W.	+				+	+		
Pectinophyton sp.					+			
Sawdonia ornata (Daws.)Hueber	+							
Kaplunella lissa Senk.	+			+	+			
Parapsilophyton balkhashense Senk.	+	+	+		+			
Psilophytites sp.				+		+	+	
Hostinella sp.	+	+	+		+	+	+	+
Lycopsides								
Lidasimophyton aff. *akkermensis* Senk.	+		+	+				
Lidasimophyton sp.	+		+		+			
Sugambrophyton sp.	+							
Protolepidodendron scharianum Krejci	+	+	+		+	+	+	+
Artschaliphyton unicum Senk.	+	+	+	+	+	+		+
Blasaria sibirica (Krysht.) Zal.	+				+	+	+	
Lycopodites aff. *runcariana* Stockm.				+				
Prelepidodendropsis kornilovae Senk.	+	+	+	+	+	+	+	+
Barrandeina elongata Senk. (in msc.)	+				+		+	
Cladoxyleans								
Calamophyton primaevum Kr. et W.	+		+		+	+		
Pseudosporochnus sp.						+		
Progymnosperms								
Rellimia aff. *thomsonii* (Daws.) Lecl. et Bonamo	+							
Incertae sedis								
Arganatiella indistinctus Stuk. et Senk.					+			

rhyniophytes, and lycopsides in assemblages of the last type, and 22 species representing algae, rhyniophytes, lycopsides, cladoxyleans, progymnosperms, and plants of uncertain systematic position are characteristic of continental deposits of the DVB and Caledonides (table). In the general taxonomic composition, the assemblages from continental formations are most similar to those from the Kurtozek Formation of the Caledonides and the Uguztau Formation of the DVB.

LANDSCAPE RECONSTRUCTIONS

It should be noted that data of paleobotany providing the basis for our reconstructions are not sufficiently reliable and complete. Methods of vegetation reconstruction suggested by Krasilov (1972) and Meyen (1988) are hardly applicable to the Devonian, especially to its middle series, because many necessary parameters are unknown. Nevertheless, we attempted

Fig. 4. Reconstructed Eifelian landscapes of the DVB and the adjacent part of the Hercynides.
The plants drawn are *Artschaliphyton* (A), *Calamophyton* (C), *Lidasimophyton* (Ld), *Protolepidodendron* (P), *Pseudosporochnus* (Ps), *Sawdonia* (S), and *Taeniocrada* (T).

to give an idea, however hypothetical, of the Eifelian vegetation in central Kazakhstan, because nothing similar had been previously discussed. The reconstruction is based on the following factors: (1) life forms of plants, (2) participation of different taxa, (3) relief character, and (4) type of enclosing deposits.

To reconstruct the relief (Fig. 4), we used data by Fedorov *et al.*, (1996) on the initial occurrence mode of the DVB volcanic rocks. According to these researchers, the most widespread landforms were volcanic plateaus and plains with separate stratovolcanoes having deluvial-proluvial aprons in slopes of some high volcanic mountains. Water basins in form of shallow lakes and streams might be widespread as well. The adjacent Hercynides incorporating deposits with marine beds were likely characterized by deltaic marshes or flooded terraces resembling those reconstructed on the basis of the Famennian vegetation the Hampshire Formation of United States (Scheckler, 1986). We think that environments in the DVB and Hercynides were more diverse and complicated than those described in this work, but no special study on the Devonian relief in Kazakhstan has yet been made. At present, we can only note that the vegetation was more diverse and abundant in the higher and dissected relief of the DVB than in the terraces, flood plains, and other places. In both cases, there was no continuous vegetation cover. The plants grew in isolated areas, which were damp or located near water basins.

Remarkably, the Eifelian plant associations from the area of the dissected relief retained rhyniophytes, which were dominant plants of the Early Devonian. They were represented by a few low upright forms with branching axes, with or without small lateral appendages (emergencies) in a form of prickles. The highest of them (up to 50 cm) were representatives of the genus *Sawdonia*. Lenticular branching shoots of *Taeniocrada* probably crept horizontally. The main components of vegetation in high mountainous plains were lycopsides representing herbaceous plants 30–50 cm high and having dichotomously branching, creeping axes, from which the aerial highly branching axes rose being thickly covered with variously forked true microleaves.

The latter promoted more active photosynthesis and gave an advantage over rhyniophytes. The abundant lycopside genus *Protolepidodendron* frequently formed monotaxial thickets. A distinctive feature of the Eifelian vegetation in the landscape under discussion was the occurrence of cladoxyleans and progymnosperms unknown from older floras. They did not form noticeable groups in biotopes and were probably represented by single specimens. Cladoxyleans were low trees. *Pseudosporochnus* and similar *Calamophyton* were at least 1.5–2 m high and had the main shoot with the crown-formed cluster of branches covered by true leaves at the top. *Rellimia*, one of the oldest but most primitive progymnosperms, was also a low tree having a hard trunk and the branch systems with, or without, flattened lateral appendages (the latter were usually spirally curved) at the top. The conductive system of cladoxyleans and progymnosperms was more developed than that of rhyniophytes and lycopsides. It was stronger and able to conduct more water. The greater height and laminate photosynthesizing organs made them more adaptable to different ecological conditions. In the Eifelian time, they grew only in high mountainous plains and near high mountains, although they also occupied other ecological niches later on. Scheckler (1986), who reconstructed the Famennian landscapes, supposed that progymnosperms, except for the Eifelian ones, preferred drier places in the flooded terraces.

The Hercynides of central Kazakhstan adjacent to the DVB probably had a weakly dissected relief with the widespread marches. From time to time, the latter were flooded by sea, and swampy environments were widespread in general. In these conditions, herbaceous lycopsides dominated by *Protolepidodendron* were not as diverse as in the DVB. Among lycopsides, there were single rhyniophytes.

The question is whether the fossil plant assemblages of the Hercynian area under discussion represent the local vegetation, or the plant detritus was transported from land into sediments of the Krykoshak and Karabastau formations. Processes of burial *in situ* and transportation of paleobotanical material were considered in detail by Krasilov (1972). Regretfully, he adduced the examples related to the post-Devonian floras. We are inclined to believe that the fossil plants of two mentioned formations are autochthonous. This is supported by the following facts. First of all, the plants under consideration grew in accumulation lowlands. The fossil material is well preserved: *Protolepidodendron schari-anum* Krejci shows dichotomously branching axes covered by small leaves dichotomous at their top; the unattached sporangia is found lying on the upper surfaces of the leaves. These features evidence for the burial conditions close to those of the plant lifetime.

CONCLUSION

The first reconstruction of the Eifelian vegetation in central Kazakhstan presented in this work is based on the paleobotanical data closely correlated with results of stratigraphic investigations. These new data were obtained as a consequence of extensive geological works supervised by A.A. Bogdanov.

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We thank all the mentioned researchers, who greatly contributed to the stratigraphy of the Devonian volcanogenic-sedimentary deposits and collected plant remains. We are also grateful to the paleobotanist S.V. Naugol'nykh from the Geological Institute of Russian Academy of Sciences, Moscow, for his creative drawing of the Eifelian landscape outlined by Yurina.

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The Berriasian Section in the Bel'bek River Basin, Southwestern Crimea

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Abstract—The Berriasian section outcropping in the right-hand side of the Bel'bek River valley (southwestern Crimea) is described. The lithological and paleontological characteristics of 10 members recognized in the section are given. The proposed new scheme of subdivision of the Berriasian in the Bel'bek section includes the following ammonite zones and beds (from the base upward): *Dalmasiceras crassicostatum* Zone (lower Berriasian), *Weberithyris moisseevi* and *Megadiceras koinautense* beds (upper Berriasian), and the overlying *Thurmanniceras pertransiensis* Zone (lower Valanginian).

Key words: *stratigraphy, Lower Cretaceous, Berriasian, Crimea*

The most complete section of Berriasian deposits in southwestern Crimea outcrops along the right-hand side of the Bel'bek River valley (Fig. 1), in ravines and scours at slopes of rises composing the southern band of the second mountainous ridge in Crimea between the Bechku Saddle (the Chernaya-Bel'bek rivers watershed) and Mt. Kaya-Tepe (the Kharu-Kacha rivers watershed). The best exposures are located in the Kabanii, Bezymyannyi, and Orekhovyi ravines (Fig. 3), where Yanin studied them in 1959, 1968, and 1990, and Baraboshkin in 1992–1997. The integral section of the Berriasian stage that is compiled using these data is characterized by several peculiar features: (1) it is rich in shallow-water facies (conglomerate, sandstone, limestone); (2) hiatuses are abundant in the section; (3) separate members are variable in thickness; and (4) abundant fossils are irregularly distributed. Despite these peculiarities, the considered Berriasian section is most complete and typical for the entire southwestern Crimea. Thus, its detailed description and paleontological characteristics are of great importance for subsequent geological investigations. The article is aimed precisely at the detailed description of the Berriasian Bel'bek section.

In addition to our own data on ammonites (Baraboshkin and Yanin, 1997; Baraboshkin in Arkad'ev *et al.*, 1997) and bivalves (Yanin and Smirnova, 1981; Yanin and Vishnevskii, 1989), the published data on the following fossil groups were used: foraminifers (Gorbachik *et al.*, 1970, 1975), corals (Kuzmicheva, 1960, 1972), gastropods (Golovinova, 1970), ammonites (Drushchits, 1956, 1960), belemnites (Kabanova, 1960), brachiopods (Smirnova, 1972, 1984; Lobacheva, 1983), and crinoids (Arendt, 1974; Klikushin, 1987, 1992). Some determinations of fossils were revised using *Atlas melovoi fauny...* edited in 1997. L.E. Vishnevskii performed the study and description of thin sections .

The Berriasian Bel'bek section was studied over a long time (Table 1). Based on paleontological studies, boundaries and ranges of different members were permanently revised and specified. For instance, the limestone member with coral bioherms, which was considered as being the Urgonian (Barremian) in age in the 1930–1940s, was interpreted to be the Valanginian at the end of the 1950s, whereas nowadays, it is referred to the upper Berriasian. Since 1956, the age of quartz conglomerate overlying the mentioned biohermal limestone was gradually changed from the late Albian to the late Berriasian.

Recently, owing to new finds and the revision of previously described ammonites, the early Hauterivian age of pudding sandstone bed was changed for the early Valanginian, and it is correlated with the lower part of the section located in the Verkhotur'e Settlement, the Kacha River area (Arkad'ev *et al.*, 1997; Baraboshkin *et al.*, 1994; Baraboshkin and Yanin, 1997).

Unfortunately, the low ammonite abundance is unfavorable for the precise dating of members I and V–X, and also for their direct correlation with the ammonite scale. This disadvantage is neutralized by finds of other guide fossils (brachiopods, rudists, and others) that permit the described members to be more or less confidently dated.

What follows below is a description of the integral section of Berriasian deposits outcropping along the right side of the Bel'bek river (from the base upward, see Fig. 2).

The underlying strata are represented by the rhythmical alternation of sandstone, siltstone, and argillite beds of the Tavriya Group (Upper Triassic–Middle

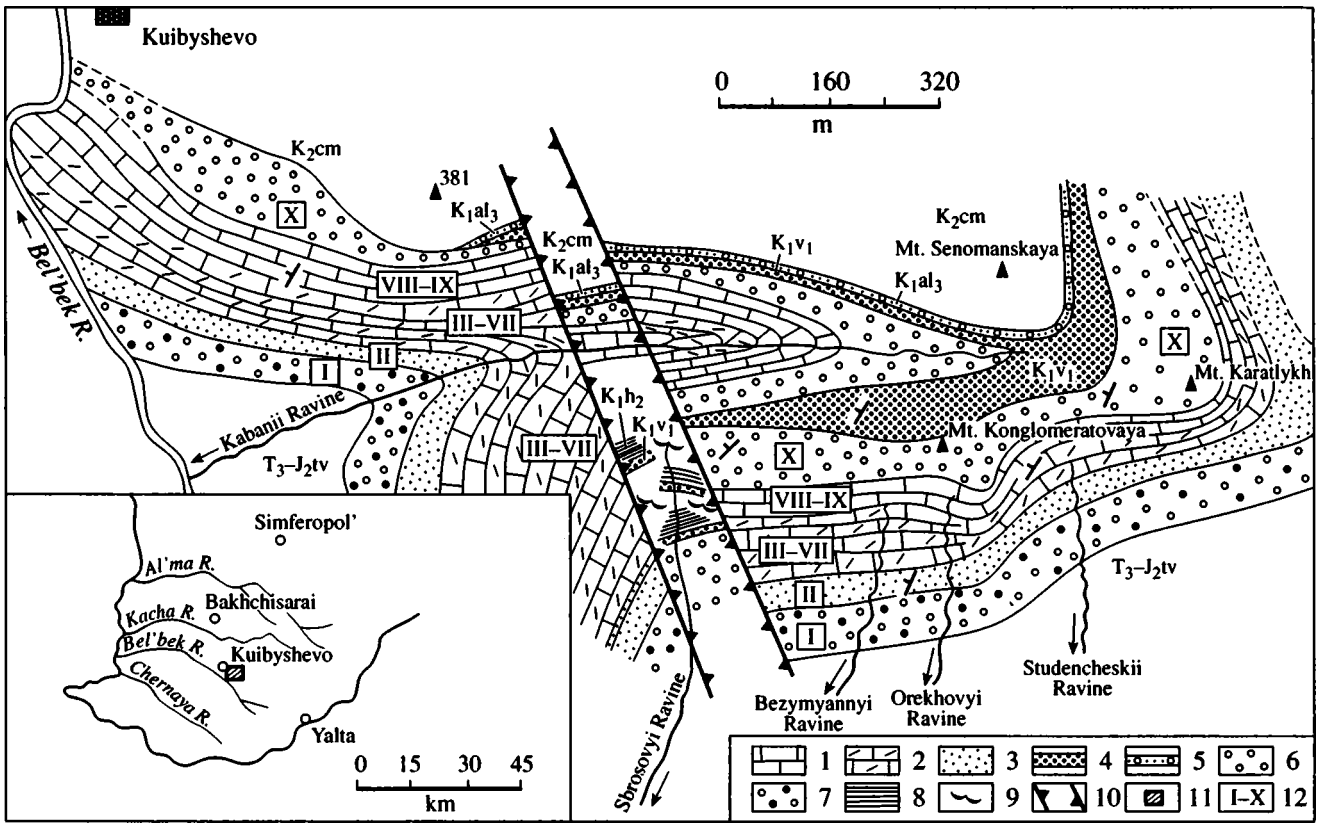

Fig. 1. Schematic distribution of Lower Cretaceous deposits in the right-hand side of the Bel'bek River valley (compiled by B.T. Yanin): (1) organogenic-detrital and biohermal limestones; (2) microphytolitic, detrital, and sandy limestones; (3) alternating loose and compact sandstones; (4) pudding sandstone; (5) glauconite sandstone with pebbles; (6) predominantly quartz conglomerate; (7) polymictic conglomerate; (8) clays with aptychuses; (9) landslide topography; (10) graben; (11) the study area; (12) members and their numbers. Indices: (K_2cm) Cenomanian Stage; (K_1al_3) upper Albian Substage; (K_1h) Hauterivian Stage; (K_1v) Valanginian Stage; (T_3–J_1tv) Tauriya Group.

Jurassic) exposed in deep scours at the southwest slope of Mt. Konglomeratovaya.

Member I (40 m) is composed of a dark to brownish gray small-pebbled polymictic conglomerate resting with the stratigraphic and structural unconformity upon the eroded surface of the Tavriya Group. Pebble dimensions vary from 25–35 (23–39%) to 35–40 mm (4–10%); less commonly, pebbles are over 50 mm (8–10%) in diameter. Sometimes, there are boulders up to 20 cm across. In the lithological aspect, pebbles mostly consist of quartz, quartzite, siltstone, and sandstone (40–46%), but also of argillite (6–18%), volcanic rocks, and metamorphic schist (5–10%).

Pebbles are well cemented by a dark-gray clayey-limy matrix of the basal type (to 15–30% of the total rock volume), locally ferruginous. The conglomerate structure is lenticular-laminated and cross-bedded. The sequence encloses interbeds of sandstone (0.2–1.2 m), siltstone, and clay. Polymictic sandstone with clayey cement is poorly sorted varying in structure from coarse- to fine-grained; siltstone is mainly quartzose and shows distinct cross bedding; clay is hydromicaceous and silty. Conglomerate contains rare coalified wood fragments. The clayey lentil in the uppermost part of the conglomerate member yielded rare foraminifers *Lenticulina* ex gr. *macra* Gorb., *Trocholina alpina* Leup., and *Epistomina caracolla* (Roem.).

Member II (14.5 m) is represented by alternating compact and slightly cemented sandstone. The member is poorly exposed in the right-hand wall of the Kabanii Ravine and at the Mt. Konglomeratovaya slopes (Fig. 3). The member begins with the basal conglomerate layer overlying the eroded surface of the conglomerate Member I. It differs from the latter by a higher abundance of gravel grains and by a more uniform composition of pebbles.

The basal conglomerate is mainly dark gray, locally brownish gray being composed in general of quartz and siliceous pebble and gravel (82%); less common are siltstone, argillite, and polymictic sandstone pebbles. The pebble size ranges from 10–25 (31%) to more than 35 mm (6%). They are irregularly cemented: in the middle part of the member, there is a compact plate (0.2 m) of variegated conglomerate underlain and overlain by the loose sandy pebblestone. In the Kabanii Ravine, this horizon is represented by a thin (0.3 m) bed

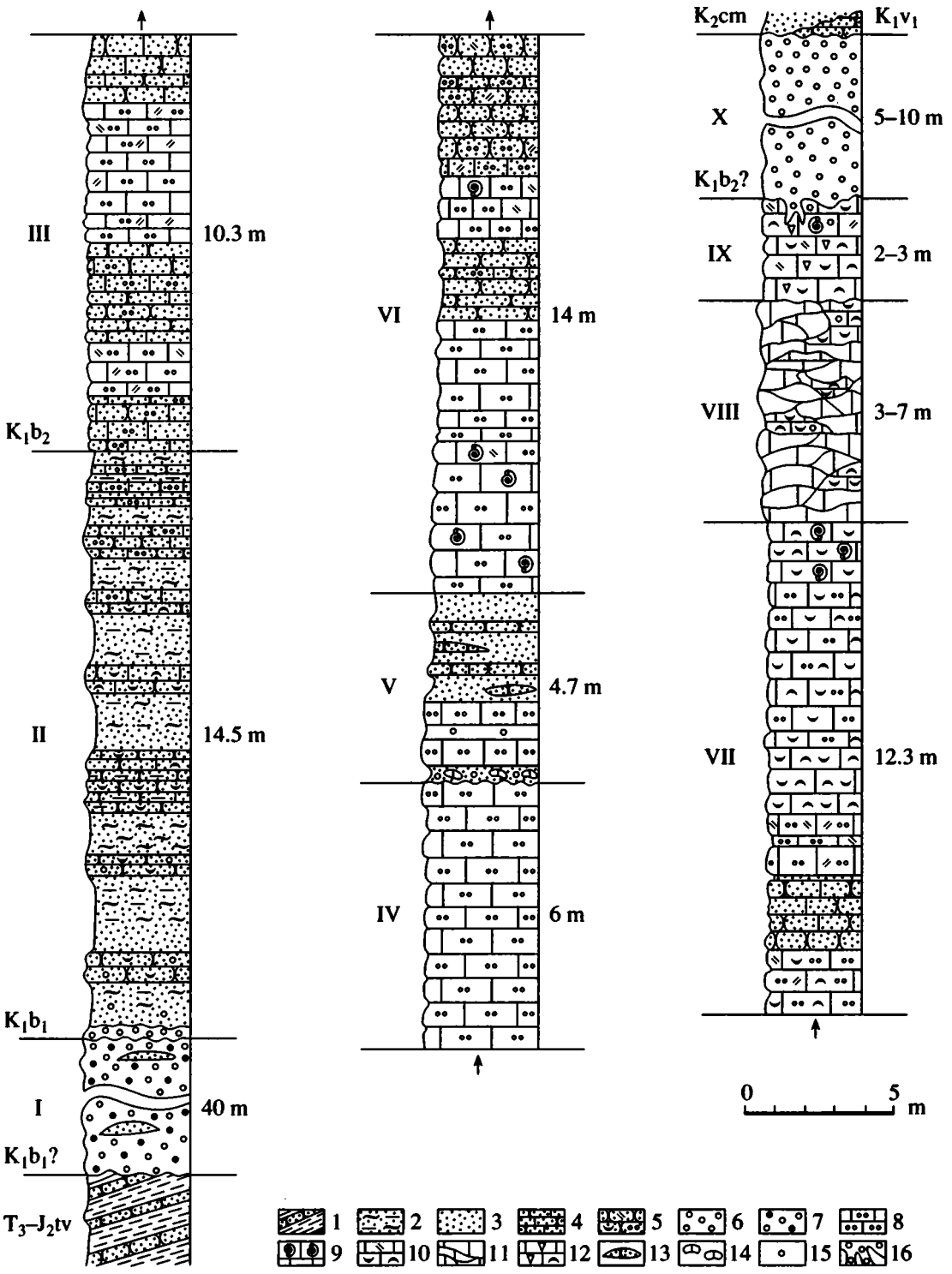

Fig. 2. The Berriasian section in the right-hand side of the Bel'bek River.

(1) alternating sandstone and argillite of the Tavriya Group; (2) clayey to silty; (3) slightly calcareous (loose); (4) highly calcareous (compact), and (5) polymictic sandstones, the latter bearing detritus and oncolites; (6) predominantly quartz and (7) polymictic conglomerates; (8) fine oncolitic; (9) coarse oncolitic; (10) organogenic-detrital; (11) biohermal (mostly coral), and (12) rudist limestones; (13) sandstone lentils; (14) limy boulders in the reworked horizon; (15) isolated pebbles; (16) neptunic dykes.

Table 1. The investigation history of Berriasian sections in the Bel'bek River valley

Stage, substage	Member (thickness, m)	Lithology	Zone, beds	Sequence	Karakash, 1907	Weber, 1937	Muratov, 1947	Drushchits, 1956	Eristavi, 1957	Drushchits *et al.*, 1958; Lychagin, 1969; Muratov, 1960, 1973; Tseisler, 1959; Shalya *et al.*, 1959	Gorbachik *et al.*, 1970; Golovinova *et al.*, 1970; Plotnikova *et al.*, 1984	Lichagin *et al.*, 1971	Gorbachik *et al.*, 1975; Yanin *et al.*, 1981	Drushchits, 1975	Kravtsov *et al.*, 1978	Aliev *et al.*, 1985; Drushchits *et al.*, 1986; Kuznetsova *et al.*, 1985; Yanin *et al.*, 1989	Bogdanova *et al.*, 1981	Nikishin *et al.*, 1983	Arkad'ev *et al.*, 1997
K_1v_1		Pudding sandstone	*Thurmanniceras pertransiens*	Karatlykh		K_1al	K_1al	$K_1al_3^2$		K_1h	K_1h_1	K_1h_1	K_1h_1			K_1h			K_1v_1
K_1b_2?	X (5–10)	Quartzose conglomerate		Upper conglomerate					K_1al		K_1v_2		K_1v_2			K_1v_2			
K_1b_2	IX (2–3)	Organogenic-detrital limestone	*Megadiceras koinautense* Beds	Kuchkino	K_1nc	K_1ur	K_1h–br	K_1br	K_1br	K_1v_{2-3}	K_1v_1	K_1v_{1-2}	K_1v_1	K_1v_1	K_1v	K_1v_1	K_1v		K_1b
	VIII (3–7)	Biohermal limestone	*Weberithyris moisseevi* Beds																
	VII (12.3)	Limestone organogenic-detrital and finely oncolitic				?													
	VI (14)	Alternating finely oncolitic and organogenic-detrital limestone and sandstone				K_1h											K_1b_2	K_1b_2	
	V (4.7)	Sandstone with limestone interbeds				?							K_1b^{3-4}	K_1b		K_1b_2			
	IV (6)	Finely oncolitic limestone																	
	III (10.3)	Alternating finely oncolitic limestone and sandstone				K_1v													
K_1b_1	II (14.5)	Alternating loose and compact sandstone	Dalmasiceras crassicostatum	Sandy-silty			K_1h		K_1v	K_1v_1	K_1b	K_1b			K_1b				
K_1b_1?	I (10)	Polymictic conglomerate		Lower conglomerate		J_3	K_1v_1		K_1v_1		$K_1b(J_3t)$		K_1b^{1-2}	J_3t		K_1b_1	K_1b_1	K_1b_1	

Note: (al) Albian, (br) Barremian, (h) Hauterivian, (v) Valanginian, (b) Berriasian, (t) Tithonian, (nc) Neocomian, (ur) Urgonian, (J_{1tv}) Tauriya Group.

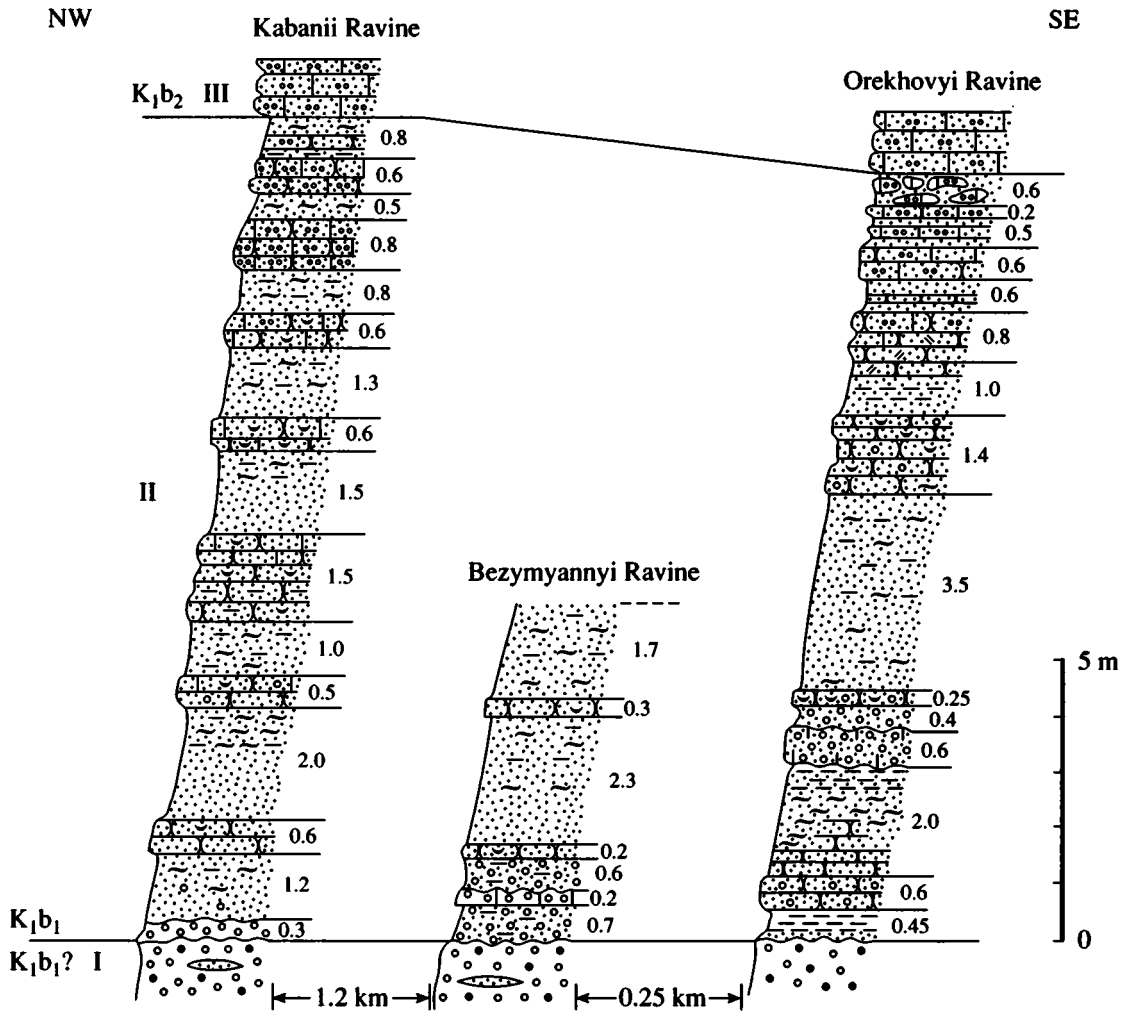

Fig. 3. Correlation of Members I-III outcropping in the Kabanii, Bezymyannyi, and Orekhovyi ravines (numbers in the right column indicate thickness of beds in meters; symbols as in Fig. 2).

of loose conglomerate with the strongly eroded roof. In its typical form, the sandstone interbeds proper are best exposed in the Kabanii Ravine. There, the member includes eight plates and compact layers of highly calcareous sandstone separated by interbeds of calcareous varieties. As a result of irregular rock weathering, the slope shows the step-like patterns (Fig. 4).

The conglomerate layer is thickest (1.5 m) in scours at the slope of Mt. Konglomeratovaya (Bezymyannyi and Orekhovyi ravines). In the Bezymyannyi Ravine, the conglomerate yielded diverse fossils: bivalves *Gervillella anceps* (Desh.), *Entolium germanicus* (Woll.), *Neithea valangiensis* (Pict. Et Camp.), *Pterotrigonia caudata* (Ag.), *Linotrigonia belbekensis* Yanin, *Sphaera belbekensis* Yanin, and *Protocardia sphaeroidea* (Forb.); brachiopods *Sellithyris uniplicarta* Smirn.; and echinoids *Toxaster* sp.

Sandstone is dirty to dark gray, fine-grained, calcareous, consisting mostly of quartz (50–70%), limestone (15–20%), and quartzite (10–15%) grains; feldspar, limonite, glauconite, zircon, tourmaline, garnet, and others minerals occur as single grains. Clasts are poorly sorted, and the grain size varies from 0.1 to 0.25 (60%), less commonly, to 0.25–0.5 mm (10%). Sometimes, gravel grains and pebbles from 2 to 15 mm across occur. Grains are usually semirounded to angular; the silt admixture is rather abundant (30–40%). Interbeds of poorly cemented sandstone contain to 40% of silty grains. Coalified plant fragments ranging from 2 to 40 mm in size are permanent components of the rocks.

Some layers contain abundant fossil remains of invertebrates represented mainly by bivalve shells. Sometimes, the latter form coquina layers or, locally, coquina pavements indicative of highly dynamic (wave-affected) sedimentation environments. Noticeable, too, is the gradual increase in the carbonate content toward the upper part of the member, where the content of both the cementing material and oncolite abundance increases. Fossils are most abundant in the lower and middle portions of the member, which

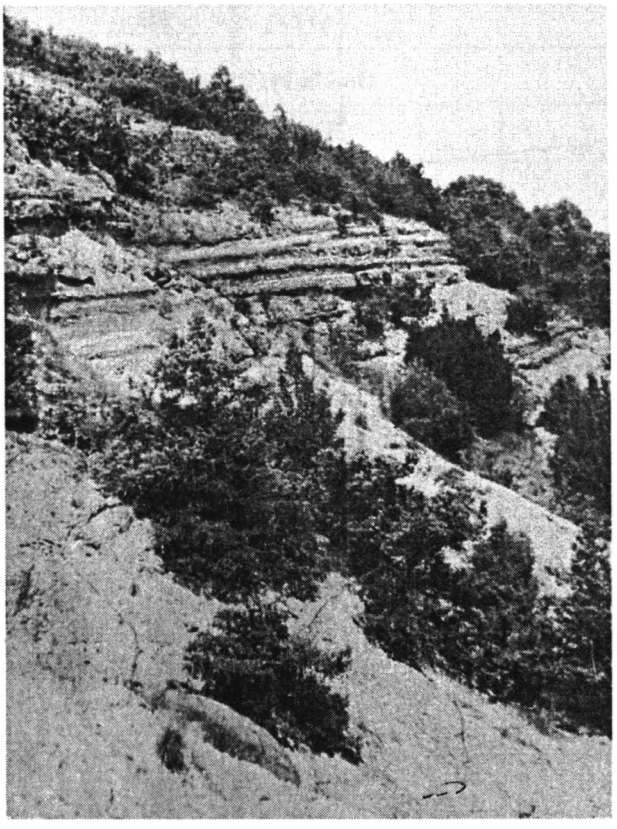

Fig. 4. The exposure of Berriasian rocks in the right-hand wall of the Kabanii Ravine (photograph by B.T. Yanin).

yielded the following assemblages: foraminifers *Rhizammina indivisa* Brady, *Triplasia emslandensis acuta* Bart. et Brand, *Gaudryina neocomica* Chal., *G. subglobosa* Ant., *Trocholina alpina* (Leup.), *Lenticulina macra* Gorb., *Citharina cristellaroides* Reuss, *Dentalina nana* Reuss, *Tristix insignia* (Reuss), *Planularia crepidularis* Roem., *Epistomina caracolla caracolla* (Roem.), *Conorboides valendisensis* (Bart. et Brand), and others; scleractinians *Montlivaltia kaufmani* Koby, *M. minima* Kusm., *M. crimea* Kusm., *M. conica* Kusm., *Paramontlivaltia valangiensis* Kusm., *Monocyclastraea alpina* (Koby), *Microsolena guttata* Koby, *Synastrea icaunensis* (d'Orb.), and *Diplosmilia taurica* Kusm.; gastropods *Scurria balaclavensis* Pchel., *Turbinopsis multicostatula* Pchel., *Ampullospira cossmanni* Pchel., *Mitriomphalus subclathratus* (d'Orb.), *Cyphosolenus vanagiensis* (Pict. et Camp.), *C. robinaldinus* (Pict.), *Perissoptera acuta* (d'Orb.), *Trochonatica submexilhoeirensis* (Pchel.), *Harpagodes jaccardi* Pict. et Camp., and *Aptyxiella belbecensis* Pchel.; bivalves *Idonearca gabrielis* (Leym.), *I. forbesi* (Pict. et Camp.), *Cosmetodon carteroni* (d'Orb.), *Gervillella anceps* (Desh.), *Gervillaria allaudiensis* (Math.), *Entolium germanicus* (Woll.), *Prohinnites renevieri* (Coq.), *Neithea valangiensis* (Pict. et Camp.), *N. simplex* Mordv., *Antiquilima dubisensis* (Pict. et Camp.), *Plagiostoma aubersonensis* (Pict. et Camp.), *Acesta longa* (Roem.), *"Lima" nicoleti* (Pict. et Camp.), *Limatula tombeckiana* (d'Orb.), *Spondylus roemeri* Desh., *Pycnodonte weberae* Yanin, *Ceratostreon minos* (Coq.), *Arcomytilus couloni* (Marcou), *Inoperna gillieroni* (Pict. et Camp.), *Myophorella loewingsonlessingi* (Renng.), *M. mordvilkoae* Yanin, *Rutitrigonia longa* (Ag.), *Ptertrigonia caudata* (Ag.), *Linotrigonia belbekensis* Yanin, *Eriphyla crimica* Yanin, *Sphaera belbekensis* Yanin, *Isocardia neocomiensis* d'Orb., *Tortarctica weberi* (Mordv.), *Protocardia peregrina* (d'Orb.), *P. sphaeroidea* (Forb.), *Integricardium deshayesianum* (Leym.), *Ptychomya robinaldina* (d'Orb.), *"Gari" valangiensis* (Pict. et Camp.), *Panopea neocomiensis* (Leym.), *Laternula agassizi* (d'Orb.), *Pholadomya gigantea* (Sow.), *Pachymya crimica* Yanin, *Goniomya archaici* (Pict. et Camp.); ammonites *Ptychophylloceras semisulcatum* (d'Orb.), *P. ptychoicum* (Quenst.), *Holcophylloceras callypso* (d'Orb.), *Biasaloceras liebigi* (Opp.), *Protetragonites tauricus* Kulj.-Vor., *Haploceras carachtheis* (Zeuschn.), *H.* ex gr. *elimatum* (Opp.), *Spiticeras orientale* Kil., *Dalmasiceras crassicostatum* (Djan.), *D.* aff. *crassicostatum* (Djan.), *D.* sp., *Malbosiceras* sp., *?Euthymiceras* sp., and *?Neocosmoceras* sp.; belemnites *Duvalia lata* (Bl.); brachiopods *Praecyclothyris gracilis* Lob., *Cyclothyris airgulensis* (Moiss.), *C. multiformis* (Roem.), *C. mutabilis* Lob., *C. minor* (Lor.), *Sulcirhynchia semenovi* (Moiss.), *Sellithyris uniplicatus* Smirn., *Simphythyris arguinensis* (Moiss.), *Tropeothyris salevensis* (Pict.), *Loriolithyris valdensis* (Lor.), *Dictyothyris spinulosa* Smirn., *Psilothyris airgulensis* (Moiss.), and *Zeillerina walkeriformis* Smirn.; crinoids *Isocrinites lissajouxi* (Lor.); echinoids *Orthopsis* cf. *repellini* Gras., *Pygopyrina uncisa* (Ag.), *Pygurus rostratus* (Ag.), *Collyropsis iheyrati* (Ooster), and *Toxaster granosus* (d'Orb.).

Member III (10.3 m) consists of alternating brownish gray fine-grained oncolitic-detrital limestone and highly calcareous brownish gray sandstone. The member is well exposed in the Kabanii Ravine and at the Mt. Konglomeratovaya slopes. Thick limestone plates are distinctly manifested in the topography. They contain abundant oncolites from 0.04 to 0.12 mm in size, which are referred to the *Osagia* group and associate with skeletal fragments of comparable size; the latter belong to different organisms: echinoids, scleractinians, bivalves, brachiopods, and foraminifers. Skeletal fragments of these invertebrates usually occur as cores of oncolites. Limestone shows an insignificant (to 5%) admixture of terrigenous material. Fine- to medium-grained, mostly quartz (to 60% of the detrital material) sandstone beds show obscure bedding and a noticeable (to 10%) content of porphyrite grains; these rocks also include oncolites and skeletal fragments. Their fine-grained matrix consists of calcite. Lenticular sandstone interbeds are up to 2.5 m thick. The member yielded the following fossils: foraminifers *Triplasia emslandensis acuta* Bart. et Brand, *Tritaxia pyramidata* Reuss, *Gaudryina neocomica* Chal., *G. subglobosa* Ant., *Len-*

ticulina macra Gorb., *Planularia crepidularis* Roem., *Conorboides valendisensis* (Bart. et Brand), *Discorbis praelongus* Gorb., and *D. crimicus* Schokh.; scleractinians *Paramontlivaltia valangiensis* Kusm., and *Monocyclastraea alpina* (Koby); gastropods *Ampullospira cossmanni* Pchel.; bivalves *Cosmetodon carteroni* (d'Orb.), *Gervillella anceps* (Desh.), *Entolium germanicus* (Woll.), *Prohinnites renevieri* (Coq.), *Neithea simplex* Mordv., *N. valangiensis* (Pict. et Camp.), *Antiquilima dubisensis* (Pict. et Camp.), *Limatula tombeckiana* (d'Orb.), *Rhynchostreon tombeckianum* (d'Orb.), *Myophorella loewingsonlessingi* (Renng.), *M. mordvilkoae* Yanin, *Eriphyla crimica* Yanin, *Tortarctica weberi* (Mordv.), and *Laternula agassizi* (d'Orb.); belemnites *Duvalia dilatata* (Bl.); brachiopods *Cyclothyris airgulensis* (Moiss.), *Weberithyris moisseevi* (Weber), *Sellithyris uniplicatus* Smirn., *Psilothyris airgulensis* (Moiss.), and *Zeillerina walkeriformis* Smirn.; crinoids *Solanocrinus infracretaceous* (Ooster) and echinoids *Toxaster granosus* (d'Orb.).

Member IV (6 m) is composed of brownish gray oncolitic limestone with obscure bedding; the thick-platy beds of the rock are exposed as distinct cliffs and scarps at the slope. In lithology, the rock is similar to limestone from Member III. The roof of the member is marked by the hard ground surface. The following fossils are characteristic of the member: foraminifers *Lenticulina macra* Gorb.; bivalves *Neithea simplex* Mordv., *N. valangiensis* (Pict. et Camp.), and *Antiquilima dubisensis* (Pict. et Camp.) associated with brachiopods *Monticlarella desori* (Pict.), *Cyclothyris airgulensis* (Moiss.), *Sulcirhynchia semenovi* (Moiss.), *Sellithyris uniplicatus* Smirn., and *Terebrataliopsis quadrata* Smirn.

Member V (4.7 m) consists of yellow-gray irregularly cemented sandstones (from loose to compact highly calcareous varieties forming lenticular interbeds and nodular aggregates). The loose sandstone varieties enclose local accumulations of abundant pebbles. The lower part of the member hosts two plates (each 0.6 m thick) of light-gray compact oncolitic limestone. In the Kabanii Ravine, the member begins with the basal pebblestone enclosing well-rounded fragments of colonial and solitary corals and bivalve shells. The pebbly layer rests on the eroded surface of underlying limestone. This reworked horizon yielded the following fossils: foraminifers *Trocholina elongata* (Leup.), *T. burlini* Gorb., *T. molesta* Gorb., *Discorbis crimicus* Schokh.; scleractinians (hundreds of specimens) *Montlivaltia kaufmani* Koby; bivalves *Entolium germanicus* (Woll.), *Prohinnites renevieri* (Coq.), *Neithea valangiensis* (Pict. et Camp.), *Acesta longa* (Roem.), *Antiquilima dubisensis* (Pict. et Camp.), *Myophorella loewingsonlessingi* (Renng.), *M. mordvilkoae* Yanin, *Rutitrigonia longa* (Ag.), *Eriphyla crimica* Yanin, *Coelopis benecki* (Boehm.), *Tortarctica weberi* (Mordv.); brachiopods *Terebrataliopsis quadrata* Smirn.

Member VI (14 m) is represented by interlayering fine-detrital oncolitic thick-platy limestone and highly calcareous sandstone with obscure bedding. Limestone is virtually similar to that from Members IV and V. The basal part of the member (4 m) corresponds to the limestone bed with larger oncolites. Sandstone is graywacke, medium-grained, massive, with the basal calcite cement fine-grained in structure. Clasts are represented by quartz, quartz sandstone, siliceous rocks, volcanics, micritic limestone; present are also fragments of bivalve shells, foraminifers, and algae. The member encloses fossils: foraminifers *Lenticulina* aff. *muensteri* (Roem.); bivalves *Prohinnites renevieri* (Coq.), *Neithea valangiensis* (Pict. et Camp.), and *Spondylus roemeri* Desh.; brachiopods *Terebrataliopsis quadrata* Smirn. and *Psilothyris airgulensis* (Moiss.).

Member VII (12.3 m) corresponds to gray thick-platy organogenic-detrital and oncolitic limestones with obscure bedding; the lower portion of the member encloses the interbed (2 m) of sandstone similar to that from the underlying member. Limestone contains fragments of bivalve, brachiopod, and foraminifer shells, scleractinians, crinoids, echinoids, and fragmented algae. Fragments of biomorphic (coral) and micritic limestone (to 30%) are associated with quartz and quartzite sandstone grains (to 5–10%). The cement is calcitic, fine-grained, basal, and interstitial. The base of the member is represented by the limestone bed (1.6 m) saturated with brachiopod shells (brachiopod horizon).

The member bears the diverse fossil assemblages: foraminifers *Tritaxia pyramidata* Reuss and *Hoeglundina* aff. *angusticostata* Amt. et Gned.; scleractinians *Thamnasteria digitata* From. and *Stylina multiseptata* Kusm.; bivalves *Prohinnites renevieri* (Coq.), *Neithea valangiensis* (Pict. et Camp.), *N. simplex* Mordv., *Plagiostoma aubersonensis* (Pict. et Camp.), *"Lima" neocomiensis* (d'Orb.), *Limatula tombeckiana* (d'Orb.), *Spondylus roemeri* Desh., *Rastellum rectangularis* (Roem.), *Ceratostreon minos* (Coq.), and *C. tuberculiferum* (Koch et Dunk.; brachiopods *Monticlarella desori* (Lrb.), *Cyclothyris airgulensis* (Moiss.), *Weberithyris moisseevi* (Weber), *Sellithyris uniplicatus* Smirn., *Terebrataliopsis quadrata* Smirn., *Advenina villersensis* (Lor.), *Zeillerina baksanensis* Smirn., and *Z. walkeriformis* Smirn.; crinoids *Apiocrinus* cf. *valangiensis* Lor. and *Isocrinites lissajouxi* (Lor.).

Member VIII (3–7 m) consists of gray limestone, organogenic-detrital, nodular, with obscure bedding, locally massive, and enclosing bioherms to 3–5 m high and 1.5–4 m wide. Bioherms are composed of dendroid and nodular coral colonies, and also of coating and nodular algae. Scleractinians are represented by abundant *Thamnastraea mammelonata* Turn., *Dermosmilia cretacica* Turn., *Placophylia grata* Burg., *Cyathophora excavata* From., *Diplocoenia octoseptata* Kusm., *D. polygonalis* Kusm., *D. decaseptata* Kusm., *Calamophyllia compressa* (d'Orb), and *Stylina pachystylina*

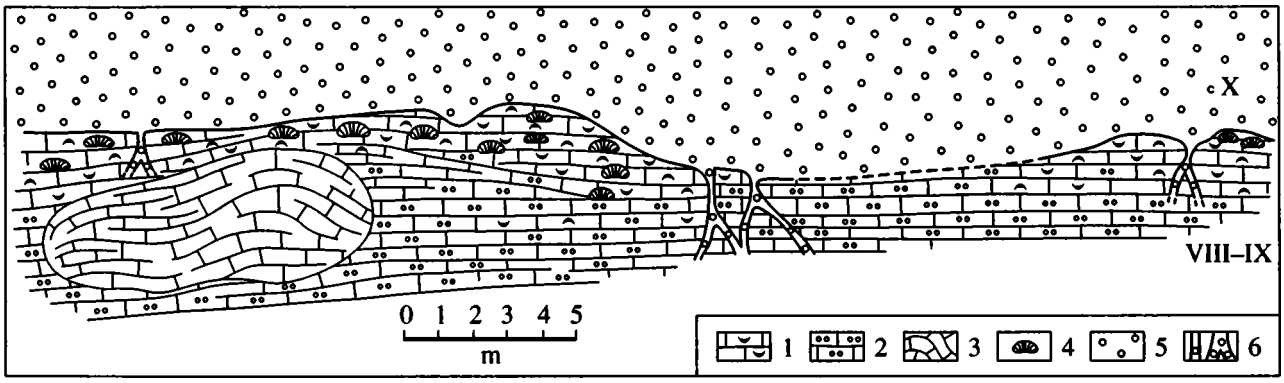

Fig. 5. Cross section of quartz conglomerate Member X resting on the eroded surface of the limestone sequence (Members VIII–IX) as it is seen in the right-hand side of the Bel'bek River valley directly above the mark 15/57 of the Kuibyshevo-Sokolinoe highway (drawing by B.T. Yanin): (1) organogenic-detrital limestone; (2) fine oncolitic limestone; (3) coral bioherm; (4) nodular colonial corals; (5) quartz conglomerate and gravelstone; (6) neptunic dykes in limestone members filled with quartz pebbles, gravel, and products of "pocket" wall destruction.

Koby. Biohermal limestones are cavernous with spaces between organogenic fragments filled with inequigranular calcite. Spaces between bioherms are filled with organogenic-detrital limestone composed of coral, sponge, bryozoan, algae, bivalve, crinoid, and echinoid fragments. Some fragments of biomorphic limestone and organic remains are covered by a thin micritic "envelope" being slightly impregnated by Fe hydroxides. Grains in the limestone are, as a rule, gravel-sized; the rock encloses fragments of biomorphic limestones, colonial corals, sponges, bryozoans, serpulids, gastropod shells, foraminifers, crinoids, and echinoids, which are mixed with terrigenous aleuritic-sandy grains (up to 10%) of quartz, quartzite sandstone, siliceous rocks, and weathered ferruginous volcanics. Sometimes, the terrigenous admixture can represent up to 40% of the rock volume. The matrix consists of inequigranular calcite.

The member encloses different organic remains: foraminifers *Trocholina molesta* Gorb.; bivalves *Prohinnites* sp., *Rastellum rectangularis* (Roem.), and *Ceratostreon* sp., *Trichites* sp., borer *Lithophaga avellana* (d'Orb.); brachiopods *Craniscus* sp., *Weberithyris moisseevi* Weber, and *Terebrataliopsis quadrata* Smirn.; echinoids *Diplocidaria* sp. and *Balanocidaris maresi* (Cott.); crinoids *"Isocrinus" arrierensis* Lor., *Apiocrinites* sp., *Burdigalocrinus maximus* Klik., *Loriolicrinus perforatus* Klik., *L. laevis* Klik., *L. asper* Klik., *Conoideocrinus conoideus* Klik., *Umbocrinus umbonatus* Klik., *Heberticrinus heberti* (Lor.), *H. irregularis* Klik., *Thiollericrinus belbekensis* Klik., *T. elongatus* Klik., and *T. torosus* Klik.

Member IX (2–3 m) corresponds to gray limestones, organogenic-detrital, with obscure bedding, locally platy and mainly composed of variably rounded fragments of colonial dendroid and nodular corals, crinoids, echinoids, *Solenopora* algae, rudists, brachiopods, and other organisms. All skeletal fragments are bored by lithophagous bivalves. Most of them are covered by a dark thin oncolitic crust with laminated or reticulate texture. Sometimes, the rocks enclose abundant clasts (10 × 10 mm in size) of fine-grained limestone derived from the eroded underlying member. In the uppermost part of the member, limestone becomes sandy and enriched in large pebbles and boulders (to 15 cm across) of variable lithology: white and smoky (rare case) quartz, massive sandstone and dark-gray limestone. Sometimes, the rock is pudding-like.

The member hosts the following fossil assemblage: foraminifers *Trocholina molesta* Gorb.; scleractinians *Stylina pachystylina* Koby and *Isastraea geometrica* Koby; gastropods *Turbinopsis multicostulata* (Pchel.); bivalves *Acesta orbignyana* (Math.) and *Ceratostreon tuberculiferum* (Koch et Dunk.); rudists *Megadiceras koinautense* Pchel., *Paradiceras* spp. and *Valletia* spp. (their shells locally form rudist banks); crinoids *Apiocrinites* cf. *valangiensis* Lor.; echinoids *Cidaris problematicus* Cott.

Member X (5–10 m) consists of a gray and brownish gray, highly ferruginous conglomerate and gravelstone with well-rounded small and medium-sized pebbles (1.4–3.0, less commonly, to 8 cm across); quartz pebbles are predominant. Other rare pebbles and gravel consist of sandstone, siltstone, argillite, quartzite, microquartzite, and siliceous rocks. Pebbles are cemented by Fe hydroxides. The member encloses lentils (to 0.5–1.0 m thick) of silty clay and cross-bedded sandstone. The conglomerate rests with a deep erosion on the underlying limestones of Members IX and VIII. In exposures along the highway near the power line, the uppermost layers of underlying Member IX show the presence of neptunic dykes penetrating to the depth of 4 m and filled with quartz conglomerate and gravelstone, as well as with angular limy fragments derived from the dyke walls (Fig. 5). No organic remains are discovered in the member. The conglomerate thickness varies from 5–10 m in the Kabanii Ravine to 70 m easterly, at Mt. Konglomeratovaya.

Table 2. Correlation of Berriasian section of the Bel'bek River valley with biostratigraphic subdivisions of the Crimea, northern Caucasus, and *standard* Mediterranean scale

Stage	Substage	Standard scale (Hoedemaeker *et al.*, 1993) Zone, subzone		Substage	North Caucasus, Urukh R. (Sei and Kalacheva, 1997) Zone, subzone		Stage	Crimea (Bogdanova *et al.*, 1981) Zone, subzone	Substage	Southwestern Crimea, Bel'bek R., this work Zone, subzone
Berriasian	Upper	*Fauriella boissieri*	*Tirnovella alpillensis*	Upper	Fauriella boissieri	Beds *Jabronella paquieri–Berriasella callisto*	Berriasian	*"Zeillerina" baksanensis* Beds	Upper	Quartzose conglomerate (Member X)
										Megadiceras koinautense Beds (Member IX)
								Symphythyris arguinensis Beds		*Weberithyris moisseevi* Beds (Members III–VIII)
			Berriasella picteti			*Riasanites rjasanensis–Riasanites angulicostatus*		*Tauricoceras crassicostatum* Beds		
			Malbosiceras paramimounum			*Euthymiceras euthymi*		*Euthymiceras* and *Neocosmoceras* Beds		
						Riasanites rjasanensisn–Spiticeras cautleyi				
	Middle	*Tirnovella occitanica*	*Dalmasiceras dalmasi*	Lower	*Tirnovella occitanica*	*Dalmasiceras crassicostatum*		*Dalmasiceras crassicostatum*	Lower	*Dalmasiceras crassicostatum* (Member II)
			Berriasella privasensis			*Tirnovella occitanica*		*Malbosiceras* ? sp. Beds		*Polymictic conglomerate* (Member I)
			Tirnovella subalpina			*Mazenoticeras malbosiforme*				
	Lower	*Berriasella jacobi*			*Berriasella jacobi*			*Pseudosubplanites ponticeus–Pseudosubplanites grandis* Zone		

In local outcrops along the highway west of the described exposures, transgressive overlapping of Members IX and VIII by lower Cenomanian glauconite sandstone is seen; eastward, in the Sbrosovyi Ravine, at Mts. Konglomeratovaya and Karatlykh, the quartz conglomerate of Member X is transgressively overlaid by the condensed sequence of lower Valanginian "pudding" sandstones. The uppermost layers of the latter outcropping in the Sbrosovyi Ravine yielded ammonites *Euphylloceras* sp., *?Protetragonites* sp., *Neolissoceras grasianum* (d'Orb.), *N.* sp., *Olcostephanus (Olcostephanus)* cf. *globosus* Spath, *Thurmanniceras* cf. *pertransiens* (Sayn), *T.* cf. *valdrumensis* (Sayn), *T.* sp., *Belbekiceras belbeki* Barabosch., *Pseudacanthodiscus crymicus* Barabosch., and others (Arkad'ev *et al.*, 1997, who also presented the detailed lithological characteristic of "pudding" sandstones).

Before going to the age substantiation of discriminated members, it should be noted that we follow the concept that the Berriasian Stage consists of two subdivisions, (Table 2) as it is well substantiated by Sakharov (1990) and in other works. Accordingly, we accept the lower and upper boundaries of the Berriasian Stage as they are suggested to be placed by the International working group (Zakharov *et al.*, 1996).

According to its stratigraphic position and paleontological characteristic, Member I evidently corresponds to the basal lower Berriasian. The uppermost conglomerate layers of this member yielded *Epistomina caracolla*, the foraminiferal species characteristic of the Berriasian-Hauterivian. In the coeval layers outcropping in one of the scours at the southern slope of Mt. Karatlykh, Kravtsov and Shalimov (1978) discovered *Myophorella loewinsonlessingi*, the bivalve species typical of the Berriasian rocks in the Crimea, northern Caucasus, and Mangyshlak region. The find of the upper Tithonian ammonite species *"Perisphinctes" transitorius* Opp. from the "lower conglomerate strata" of the Bel'bek River basin (Weber, 1937) seems doubtful and needs confirmation.

Finds of ammonites *Dalmasiceras crassicostatum* (Djan.), *D.* aff. *crassicostatum* (Djan.), *D.* sp., *Malbosiceras* sp, and others in rocks of Member II allow it to be confidently referred to the lower Berriasian *Dalmasiceras crassicostatum* Zone. Accompanying ammonites *Ptychophylloceras semisulcatum, P. pty-*

choicum, Holcophylloceras callypso, Biasaloceras liebigi, Protetragonites tauricus, Haploceras carachtheis, H. ex gr. *elimatum*, and *Spiticeras orientale* are also characteristic of the lower Berriasian of the Crimea (Bogdanova *et al.*, 1984). Representatives of *Euthymiceras* sp. and *Neocosmoceras* sp. are also mentioned from this section (Arkad'ev *et al.*, 1997), but, in our opinion, the poor preservation of relevant specimens does not allow their generic affinity to be confidently established and, thus, the position of the lower-upper Berriasian boundary in the section appears to be problematic. In addition, ammonites of these genera first appear in the uppermost lower Berriasian layers together with *Dalmasiceras* representatives in the sections of southeastern France (Le Hegarat, 1973) and Balkans (Nikolov, 1982).

We refer all overlying members of the considered section to the upper Berriasian. Based on the discovered peculiar assemblage of brachiopods, Members III–VIII are united into the *Weberithyris moisseevi* Beds. This species is recorded in Members III, VI, and VIII. It is also characteristic of the organogenic-detrital and biohermal limestone sequence in central Crimea (Mezhgor'e Settlement, Mt. Baksan).

On the basis of rudist finds and its stratigraphic position, Member IX, corresponding in the Bel'bek River basin to the uppermost layers of the limestone sequence, is referred to the upper Berriasian. In the Kabanii Ravine, it yielded *Megadiceras koinautense* (Yanin, 1989), the endemic rudist species earlier recovered from the upper part of the limestone sequence of the Koinaut Rock in the Burul'cha River valley (Pchelintsev, 1959). The same locality yielded *Heterodiceras crimicus* Yanin (Yanin, 1989) very similar to *H. luci* (Defr.) from the Berriasian (Infravalanginian) of the Mont Saleve area of southeastern France (Joukowsky and Favre, 1913). Previously, this form was described from the limestone member in the Zuya River valley as *H. luci* (Yanin, 1958). In the Grande Varappe section of the Mont Saleve area, this species was detected in the lower part of the limestone succession. On this basis, the *Heterodiceras luci* Beds were distinguished. Unfortunately, the Berriasian limestone and marl of the Mont Saleve section are barren of ammonites, and therefore it is impossible to correlate these beds with the standard scale.

In addition, the upper layers of the limestone sequence of central Crimea (Beshterek and Zuya rivers) show coexistence of *Heterodiceras crimicus* with large-shelled gastropods of the *Leviathania* genus (Golovinova, 1982), which were later attributed to the *Tuberleviathania* genus and new species *T. beschterekensis* (Golovinova and Korotkov, 1986). Representatives of this genus are widespread in the Lower Cretaceous carbonate facies of southeastern France, Spain, and Portugal. For instance, in the Mont Saleve area, the *Natica leviathan* Beds are recognized in the Berriasian section directly above the *"Heterodiceras luci* Beds" (Joukowsky and Favre, 1913), whereas in the Castril and Mariola Mountains of southern Spain (Alicante), this form named as *Ampillina leviathan* was found together with *Calpionellopsis simplex* (Foucault, 1974). In Portugal (Estremadura), species *Leviathania leviathan* occurs in the middle and upper parts of the upper Berriasian and in the lower Valanginian (Peybernes *et al.*, 1975). According to Remane (1989), *Calpionellopsis simplex* is the index species of the lower subzone of the standard *Calpionellopsis simplex* (D) Zone established in the Vaucaunt depression (southern France). The subzone is correlated with the lower (*paramimmounum*) ammonite zone of the upper Berriasian *boisseri* Zone (Le Hegarat, 1971).

Based on these indirect data, we assume that the upper part of the reefal and organogenic-detrital limestone sequence in central Crimea enclosing rudists *Megadiceras koinautense, Heterodiceras crimicus* and leviathanias *Tuberleviathania beschterekensis* can be referred to the upper Berriasian (Table 2), as in two earlier works (Bogdanova *et al.*, 1981; Lobacheva, 1983), although some researchers dated these rocks back to the Valanginian (Drushchits *et al.*, 1956; Drushchits, 1960; Lysenko *et al.*, 1979). Inasmuch as we consider the member of organogenic-detrital limestones of the Bel'bek section (Member IX) as representing a stratigraphic analogue of the uppermost layers of the limestone sequence in central Crimea that was formed in the common basin and under similar conditions, we believe that the member in question can be referred in the study area to the upper Berriasian *Megadiceras koinautense* Beds.

As for Member X, it is conditionally placed in the upper Berriasian. It is hardly probable that rocks of the member correspond to the basal Valanginian. This is evident from the following consideration. As was mentioned, the uppermost layers of the overlying "pudding" conglomerate and sandstone sequence contain the ammonite assemblage peculiar of the *Thurmanniceras pertransiens* Zone (Arkad'ev *et al.*, 1997), the second one from the base of the lower Valanginian zonation (Busnardo *et al.*, 1979). The lower portion of these "pudding" rocks can be correlated with the basal part of another section similar in lithology and exposed along the Kacha River, where the lower Valanginian *Kilianella otopeta* Zone is also present (Baraboschkin and Mikhailova, 1994; Baraboschkin, 1997; Baraboschkin and Yanin, 1997). Consequently, the conglomerate Member X cannot be younger than the late Berriasian, because the "pudding" rocks themselves are composed of the material derived from reworked underlying conglomerates.

CONCLUSION

(1) The Berriasian deposits of the Bel'bek section are divided into 10 lithological members, for which the lithological-paleontological characteristic is given.

(2) The following biostratigraphic units are defined in the studied Berriasian section: the lower Berriasian

Dalmasiceras crassicostatum Zone (Member II) and the upper Berriasian *Weberithyris moisseevi* (Members III–VIII) and *Megadiceras koinautense* (Member IX) beds. Lower polymictic conglomerate strata (Member I) are arbitrarily referred to the lower substage, and the quartz conglomerate beds of Member X, to the upper substage. The latter are overlain by the pudding sandstone of the lower Valanginian *Thurmanniceras pertransiens* Zone.

(3) The distinguished units of the Berriasian section of southwestern and central Crimea are correlated with coeval subdivisions known in southeastern France and Spain, where sediments represent the shallow-water carbonate-platform facies.

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The Miocene Beshcheul Horizon of West Siberia: Stratigraphy, Paleobotany, and Paleomagnetism

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Abstract—The mid-Miocene in West Siberia was marked by significant environmental changes related to development of the river system in response to the uplift of the mountainous framing structures. This resulted in formation of the lateral succession of several alluvial formations and in the landscape reorganization that affected, in particular, vegetation (the gradual replacement of the mesophilic Turgai flora by the post-Turgai forest-steppe formations). These changes are recorded in the Beshcheul Horizon representing the boundary stratigraphic unit of the region, above which the Oligocene to lower-middle Miocene succession of humid gray-colored deposits of the West Siberian platform cover is replaced by a new semiarid sedimentary sequence accumulated in the second half of the Miocene-Pliocene time. In this work, we consider reference sections of the horizon (Irtysh and Ob' river basins) and their biostratigraphic and paleomagnetic characteristics. On the basis of summarizes paleobotanical data collected over many years, the evolution of the post-Turgai flora and its most important stages are outlined. The composite magnetostratigraphic section of the horizon, with five major zones of normal and reversed polarities, is also compiled. Based on the available data, the stratigraphic range of the horizon is assumed to be corresponding to the Burdigalian-Serravalian.

Key words: *stratigraphy, paleocarpology, palynology, paleomagnetism, lower-middle Miocene, Beshcheul Horizon, formation, floral level, magnetic zone, West Siberia*

INTRODUCTION

The paper continues a series of publications aimed at the perfection of Neogene stratigraphic scales of West Siberia and based on the results of multidisciplinary study. In the earlier works, we dealt with the Neogene magnetostratigraphic reference sections in the middle part of the Ob'-Irtysh interfluve area (Gnibidenko *et al.*, 1989), Omsk-Irtysh (Gnibidenko, 1990; Gnibidenko *et al.*, 1991) and Tara-Irtysh (Gnibidenko *et al.*, 1996; Martynov *et al.*, 1997) regions, which were compiled using data on natural exposures and drilling results obtained in the course of geological mapping. This article presents data on the Beshcheul, or Petropavlovsk (Martynov, 1965) regional horizon, above which the Oligocene to lower-middle Miocene succession of humid gray-colored deposits of the West Siberian platform cover is replaced by a new semiarid sedimentary sequence accumulated in the second half of the Miocene–Pliocene time. It should be emphasized that Cenozoic continental multistage successions of the West Siberian plate show a complex internal vertical and lateral stratification related to long-term migrations or progradation of alluvial, alluvial-lacustrine, and lacustrine-boggy sediments over the spacious, almost flat plain (Martynov, 1969, 1980). Such a structure of continental deposits in the vast West Siberian accumulative plain differs them from sequences similar in lithology but filling deep intermountane troughs such as, for instance, the Zaisan and other depressions.

REGIONAL DATA

The stratigraphic study of Neogene deposits in West Siberia lasted already over a century. As early as in the 1960s, it became possible to compile the correlation schemes of Neogene deposits for many areas of the plain and to subdivide the complex continental sedimentary successions into the relatively short-ranging stratigraphic units–formation and horizons. In 1967, the stratigraphic scheme was modified to include the Abrosimovka and Beshcheul horizons. At that time, the former was referred to the upper Oligocene, whereas the latter, to the lower Oligocene (mainly because of a rather sharp change in lithology and composition of spore-pollen assemblages) The Beshcheul Horizon united several coeval alluvial and alluvial-lacustrine formations distinguished in the course of geological mapping both in the marginal and central areas of the plain. For instance, the Shiderty and Krutishka formations of ancient river valleys, which are composed of inequigranular, frequently ferruginous sand with gravel and pebbles in their lower parts and of mottled to whitish clayey silt in the upper parts, were distinguished

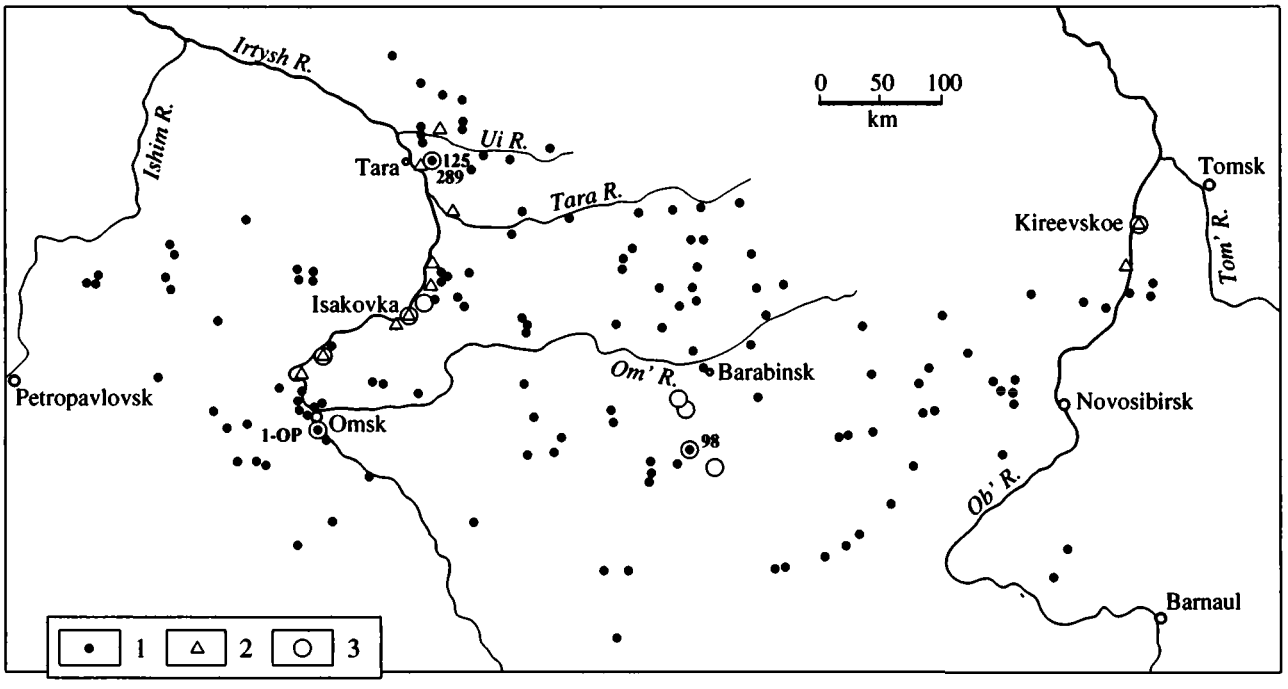

Fig. 1. Location of the studied sections of the Beshcheul Horizon: (1) borehole; (2) exposure; (3) paleomagnetic section.

along the southern margin of the West Siberian plain adjacent to Kazakhstan and in the Altai piedmonts. Farther toward the center of the plain, the Petropavlovsk, Beshcheul, Kas'kovka, and Suzun formations of sandy and sandy-silty-clayey lithology were mapped. The Bolotnoe and coeval Tagan formations of alluvial inequigranular sand bearing gravel, pebbles, and coal detritus, which was most likely transported by the paleo-Tom' river from the Kuznetsk basin, were established in the Tom'-Irtysh region located in the southeastern part of the plain. The Tagan Formation was found to be widespread northerly, in the Chulym River basin. In the Yenisei region of West Siberia, this stratigraphic level is represented by the Kirnaeva Formation composed of similar coarse-grained alluvium. It should be noted that the coeval deposits in the neighboring Turgai region are represented by the Kushuk Beds and other alluvial units accumulated in similar sedimentation settings.

The analysis of regional materials revealed that the considered lateral succession of stratigraphic units represents one horizon that reflects a sedimentation stage corresponding to the activation period of erosional and accumulative processes. In the absence of representative faunal assemblages, the age of the formations listed above was determined on the basis of their correlation with stratigraphic units similar in lithology and known in adjacent regions, and also by using paleobotanical data on both the relatively rare macrofloral remains (impressions of leaves and others) and abundant fossil spore-pollen and carpological faunas from many localities. In the latitudinal band stretching between the Irtysh and Ob' rivers, there are over 150 localities with the fossil seed flora characteristic of the Beshcheul Horizon (Fig. 1). Below, we consider reference sections of the horizon.

The Petropavlovsk area (near the TV tower, the right-hand Ishim riverside). When defining the considered horizon as an autonomous stratigraphic unit of a regional rank, the natural sections of Neogene deposits along the Irtysh, Ishim, and Ob' rivers were considered as the reference ones. The seemingly most complete section located at the Ishim River bank near Petropavlovsk and described in works by A.G. Ber, V.V. Lavrov, B.E Antypko, and B.A. Martynov was proposed to be the stratotype of the horizon. Deposits outcropping in this locality just above the water line correspond to the Abrosimovka Horizon composed of bedded sandy and silty clay rich in plant detritus, coalified wood chips and fragments, and also containing carpological assemblages; their apparent thickness is 7 m (Martynov and Nikitin, 1968). The base of the overlying sequence (Petropavlovsk Formation) corresponds to a member of alluvial cross-bedded sand (6–7 m), gravelly in the lower part and locally cemented by the ferruginous material. In neighboring areas of Kazakhstan, similar sands and sandstones are referred to the Shiderty Formation; by lithology, they are also similar to the Kushuk Beds of the Turgai region. These beds enclose remains of the Burdigalian mammals (*Homphoterium*) and early Miocene small-leaved flora. The middle part of the section (about 6 m) consists of bedded sandy clay

members alternating with beds of fine-laminated clay, silt, and fine-grained sand often bearing fine and dispersed plant detritus. Unfortunately, these layers in the Petropavlovsk exposure are barren of fossil fruits and seeds, but they contain diverse spore-pollen assemblages. According to data by G.M. Bratseva, K.A. Merkulova, and L.A. Panova, the pollen spectrum of these sediments is dominated by seed plants, most preponderant amid which are *Alnus* and diverse herbaceous forms (*Sparganium, Potamogeton*, Poaceae, Polygonaceae, Asteraceae). Simultaneously, pollen of broad-leaved plants (*Juglans, Pterocarya, Carpinus, Corylus, Quercus sibirica* Pan., *Ulmus*) continue to occur. Pollen of coniferous is dominated by Pinaceae, whereas that of Taxodiaceae occur as single grains; spores are mainly represented by Polypodiaceae. Leaf impressions also present here are, unfortunately, of a low stratigraphic value: *Arundo goeppertii* Heer, *Phragmites oeningensis* (A. Br.) Heer, *Salix lingulata* Goepp., *Populus balsamoides* Goepp., *Pterocarya castanaefolia* Goepp., and others (determinations by A.N. Krishtofovich and I.V. Vasil'ev). It should be noted that these layers outcropping in the ravine near the Borki Settlement, 5 km downstream of the Ishim River, yielded diverse seed flora, similar to that from the section near the village of Isakovka in the Irtysh River valley. The upper part of the section, including the bleached bedded silts with the quartz sand interbed that yields fish and small mammal bones in the lower portion (6 m), is defined as the autonomous Ishim Formation. These layers are overlain by a sequence of green-gray fine clay (10–12 m) enclosing calcareous concretions and referred to the upper Miocene–lower Pliocene succession of semiarid sediments. The study of Neogene sections in the Ishim River valley is particularly important, because they can be directly correlated with coeval sediments of the Turgai and northern Aral regions.

The village of Isakovka (the right-hand Irtysh riverside). The section is considered as another possible stratotype of the horizon. In 1947, V.A. Nikolaev united peculiar layers outcropping in this locality into the "Beshcheul Formation" using the unpublished data on the section located near the Beshcheul Settlement. At the Interregional stratigraphic meeting held in 1967, when, after a bitter dispute, it was decided to define the Beshcheul Horizon as an autonomous stratigraphic unit, the section exposed in the right-hand Irtysh riverside, 3.5 km downstream of the village of Isakovka, and well studied in 1963 (Nikitin, 1978) was proposed to be the horizon stratotype. In this section, the lower part of the Beshcheul Formation, whose base is not exposed, but was recovered in the prospecting pit, corresponds to the 10–12-m-thick alluvial member of fine-grained sand with wood fragments and small-sized plant detritus. The sand shows signs of intraformational reworking and encloses clayey pebbles that are particularly abundant near the base (Fig. 2). Higher, there is the 10–11-m-thick member of alternating silty sand, silt, and silty clay, which encloses accumulations of a fine plant detritus. The section of the Beshcheul Formation is crowned with two closely spaced horizons of buried soil exposed along the entire steep slope and downstream. The soil is overlain by a member of distinctly eroded, brownish (and bleached when dry) clayey silt (3 m). The upper part of the section (14 m) consists of gray polymictic micaceous sand and overlying clay with calcareous concretions; it is referred to the Miocene–Pliocene Pavlodar Formation (s. l.), or to the Nizhnyaya Il'inka and Novaya Stanitsa beds of the latter (Martynov, 1967).

The lower boundary of the Beshcheul Formation is not exposed in the considered section; however, the prospecting pit at the scarp base under basal layers of the formation recovered the horizontally bedded member with an apparent thickness of 1 m. The member is composed of clay with coalified wood fragments, silt with the clayey interbed, and sand saturated with plant detritus; the sediments are so highly inundated that it was impossible to recover the brown coal seam, whose presence is evident from coal debris squeezed by landslides. These deposits probably represent the upper layers of the Abrosimovka Formation, because they contain the carpological assemblage close to that characteristic of the Vasyugan stratigraphic level (Nikitin, 1984).

The carpological assemblage from the Beshcheul Formation includes approximately 300 species of fossil plants established here by now and representing flora of the Isakovka level. The Isakovka flora is highly dominated by seeds and fruits of aqueous-boggy herbaceous species. Most characteristic amid them are *Sparganium besczeulicum* Dorof., *S. juzepczukianum* Dorof., *Potamogeton erosus* P. Nikit., *P. corticosus* P. Nikit., *P. pectinatus* L., *P. besczeulicus* Dorof., *P. tertiatus* Dorof., *Caulinia irtyshensis* Dorof., *C. reticulata* Dorof., *Najas major* All., *Caldesia cylindrica* Dorof., *Stratiotes intermedius* Chandl., *Butomus umbellatus* L., *Carex flagellata* C. et E. Reid, *C. paucifloraeformis* P. Nikit., *Scirpus nikitinii* G. Bal., *S. longispermus* Dorof., *S. kipianiae* V.P. Nikit., *S. tertiarius* Dorof., *S. palibini* P. Nikit., *Acorus tertiarius* V.P. Nikit., *Epipremnum crassum* C. et Reid, *Lemnospermum pistiforme* V.P. Nikit., *Irtyshenia tenuicostata* Dorof., *Palaeoeuryale sukaczevi* Dorof., and *Myriophyllum pulchellum* Dorof. Present are also meadow-steppe herbaceous species: *Rumex, Polygonum, Cleome rugosa* Dorof., *Potentilla pliocenica* C. et Reid, *Valeriana tertiaria* V.P. Nikit., *Trichosanthes fragilis* C. et Reid, and also some species of Poaceae, Chenopodiaceae, Caryophyllaceae, Polygonaceae Brassicaceae, Apiaceae, Asteraceae, and other families. Remains of arboreal plants are not diverse and abundant; most common are *Morus tertiaria* Dorof. accompanied by relatively rare *Humulus, Comptonia, Betula, Alnus, Liriodendron* and *Vitus* forms; even less abundant are coniferous (*Picea, Pinus*). Genera, whose representatives occur in West Siberia until now, are as abundant as 53–55% of the total assemblage.

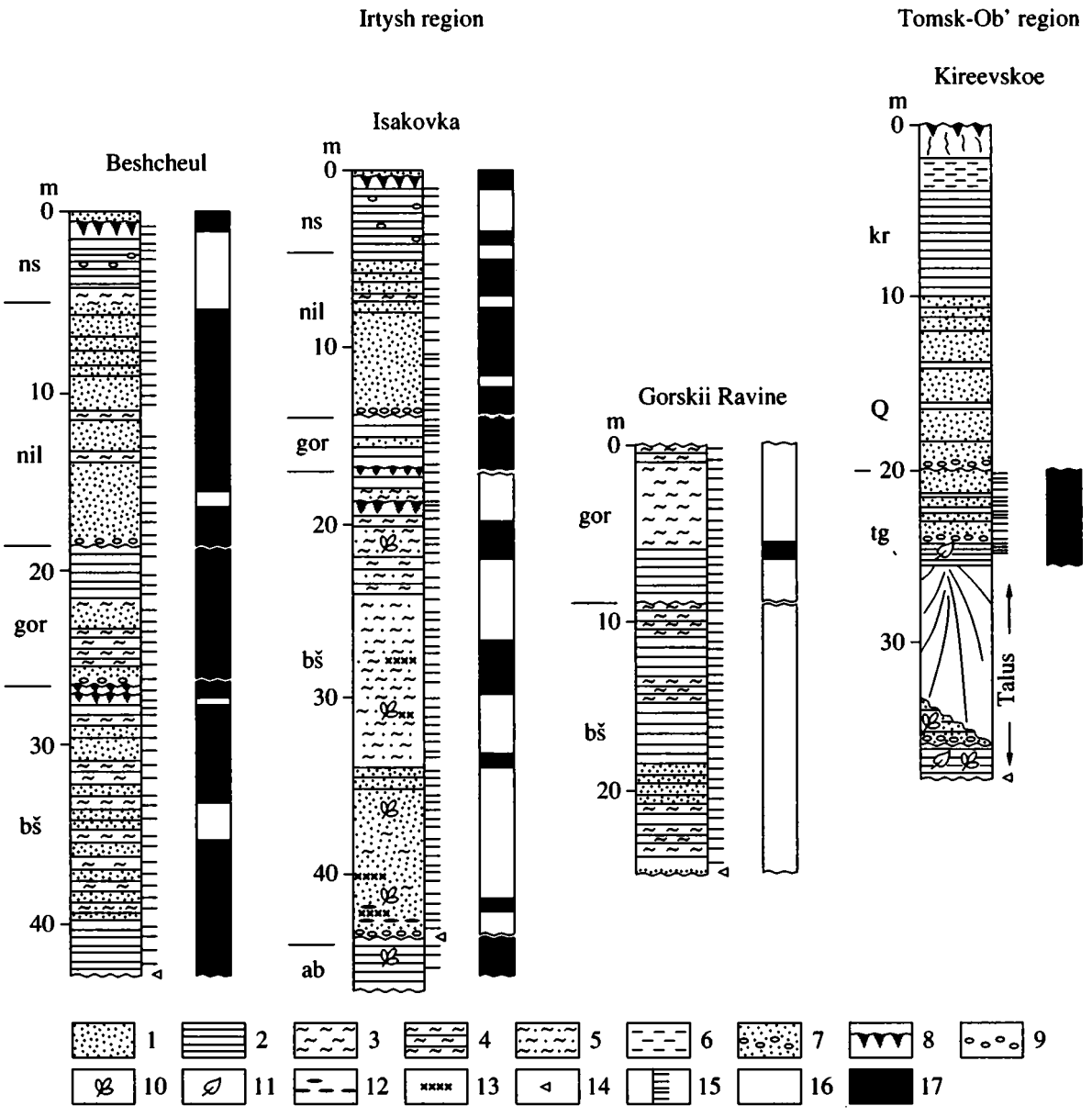

Fig. 2. Lithological-magnetostratigraphic reference sections of the Beshcheul Horizon:
(1) sand; (2) clay; (3) silt; (4) alternating clay and silt; (5) alternating sand and silt; (6) sandy loam; (7) pebblestone; (8) buried soil; (9) concretions; (10) seed assemblages; (11) leaf impressions; (12) fragments of coalified wood; (13) plant detritus; (14) water line; (15) sampling levels; (16) reversed polarity; (17) normal polarity;
(ns) Novaya Stanitsa Beds; (nil) Nizhnyaya Il'inka Beds; (gor) Gorskii Beds; (bš) Beshcheul Formation; (tg) Tagan Formation; (kr) Kireevskoe Formation; (ab) Abrosimovka Formation; (Q) Quaternary sediments.

As was previously noted (Nikitin, 1984, 1988), the carpological assemblages of the Isakovka floral level are sufficiently widespread over the plain (more than 50 localities) and characterize the post-Turgai stage of the West Siberian flora development and the forest-steppe landscape that existed under the temperate semihumid climate (substantially milder as compared with the modern one). Spore-pollen assemblages from the section located near the village of Isakovka indicate the same environments (Merkulova, 1966, 1971; Aleksandrova, 1969). Experts in palynology note that spectra from these deposits are almost devoid of Taxodiaceae, whereas Juglandaceae, *Quercus, Carpinus*, and *Nissa* forms of the Turgai time are represented by single pollen grains. Pollen of arboreal plants is dominated by boreal *Alnus, Betula*, and *Ulmus* species; the noticeable role (in some spectra up to 35–44%) is characteristic of herbaceous species representing Sparganiaceae, Poaceae, Chenopodiaceae, Asteraceae, Apiaceae, Ranunculaceae, Cyperaceae, Polygonaceae, and other families.

The section under consideration was first subjected to the paleomagnetic study that revealed magnetic characteristics of all its beds and formations (Fig. 2). The 27-m-thick sequence bearing the Isakovka flora turned out to be mostly of the reversed polarity, with four thin intervals of normally magnetized rocks (the magnetic zone R).

The Beshcheul Settlement (the right-hand Irtysh riverside, 5 km downstream from the village of Isakovka). In the section located near the Beshcheul Settlement, rocks of the Beshcheul Formation about 20 m thick are represented by different (probably, by lacustrine in most cases) facies, namely by brown clay with "banded" sand and silt. Carpological assemblages have not been discovered in this section. The bed-by-bed study of spore-pollen assemblages at this site (Volkova and Panova, 1975) showed that they are slightly richer than the section near the village of Isakovka, in terms of pollen of arboreal broad-leaved plants, especially in that of Taxodiaceae (0.4 to 4%); the herbaceous pollen is relatively less abundant (20–25%) in this section. The rock bedding features and palynological data allow us to suggest that basal layers of the Beshcheul Formation are somewhat older near the village of Isakovka than neat the Beshcheul Settlement. Nonetheless, the time gap between floras from the Abrosimovka Formation and the Isakovka level is distinct: the content of herbaceous pollen in the former does not exceed 5%, even in its upper part.

Like in the previous section, all rocks from the Beshcheul Settlement section were studied with respect to paleomagnetic characteristics. According to the latter, the studied portion of the Beshcheul Formation (18 m) is normally magnetized (the magnetic zone N) and spans only one interval of reversed magnetization (Fig. 2).

Isolated exposures of the Beshcheul Formation are investigated in the right-hand Irtysh riverside between the Chernoluch'e Settlement (north of Omsk) and Yekaterininskoe Settlement (south of Tara), and also further along the Ui River, the right tributary of the Irtysh River. In the section near the Gorskii Log Settlement, the Beshcheul Formation (16 m) corresponds to the magnetic zone with the reversed polarity (the magnetic zone R, see Fig. 2).

Several sections (Chernoluch'e, Lezhanka, Kartashovo, and others) yielded representatives of the Isakovka flora, whereas those near the Yekaterininskoe Settlement and in the Ui River areas bear the Kas'kovka flora of the same horizon (see below). In addition, we studied in detail the spore-pollen assemblages from the Isakovka floral level in some other localities. According to the data obtained, these palynological assemblages also correspond to the post-Turgai type of vegetation characterized by development of forest-steppe plant groupings with *Alnus, Betula, Ulmus*, and abundant Polypodiaceae; we also noted an increased role of herbaceous formations: Sparganiaceae, Poaceae, Chenopodiaceae, Apiaceae, Asteraceae, and others (Martynov and Merkulova, 1965). Faunal remains from the Beshcheul Formation are represented by impressions of fish skeletons and also by molluscan shells.

In fact, every section can be represented by different facies (fluvial, floodplain, lacustrine) of horizontally and cross-bedded rocks. Similarly, the diverse lithology can be also observed in borehole sections, where the Beshcheul Formation is sandwiched between coaliferous deposits of the Abrosimovka Horizon and the clayey sequence of the Tavolzhan Formation. The floral assemblages also show a slight variability, and more ancient representatives peculiar of the horizon may appear in some sections.

Sections of the Ob'-Irtysh interfluve area. In the eastern part of the study area, the Beshcheul level coincides with the alluvial Kas'kovka Formation (its stratotype corresponds to the interval of 87.5–121.5 m in the Borehole 14 near the Kas'kovka Settlement, north of Chulym) composed of fine-grained well-sorted sand in the lower part and of fine-grained, silty, and clayey sand in the upper one. Interlayers of plant detritus occur throughout the entire section. The formation rests on deeply eroded strata and is overlaid by the Miocene–Pliocene and Quaternary sequences.

The seed assemblages of the Kas'kovka floral level (Nikitin, 1965, 1984, 1988) are established in the stratotype section at depth levels of 88 and 100 m being also known in many localities of the study area (over 70 localities). They are generally similar to the Isakovka floral assemblages, although differ from the latter in a slightly lower content of endemic species (in average 49%) and in a higher share of the relict Turgai species (still present are *Metasequoia, Taxodium, Potamogeton rotundus* V.P. Nikit., *Scirpus ruthenicus* V.P. Nikit., *Betula* ex sect. *Palaeocostata, Thalictrum lanceolatum* V.P. Nikit., *Liriodendron geminatum* Kirchh., *Staphylea*, probably, *Humularia, Poliothirsis, Diclidocarya sibirica* P. Nikit., and *Nissa*); simultaneously, the Kas'kovka flora is characterized by presence of *Potamogeton trichoides* Cham. et Schlecht., *Hydrocharis morsus-ranae* L., *Boehmeria dubia* V.P. Nikit., *Urtica pliocenica* Dorof., *Chenopodium polyspermum* L., *Thalictrum simplex* L., *Diervilla tertiaria* V.P. Nikit. and some other plants typical of the post-Turgai or even modern flora. The spore-pollen assemblages from layers with the Kas'kovka flora are dominated by pollen of seed plants, mostly of *Alnus* (to 25%), *Betula* (to 10%), *Ulmus, Salix, Juglans, Carya*, and other genera. Herbaceous pollen is sufficiently diverse, but its abundance does not exceed 15–20%; the share of coniferous pollen represented by *Picea, Pinus, Tsuga, Abies*, and single Taxodiaceae amounts to 8–10%; Polypodiaceae species are preponderant amid spores (Merkulova, 1971; Volkova and Panova, 1975). Judging from the paleobotanical data, the prevailing landscapes in the southern areas of West Siberia were represented in the

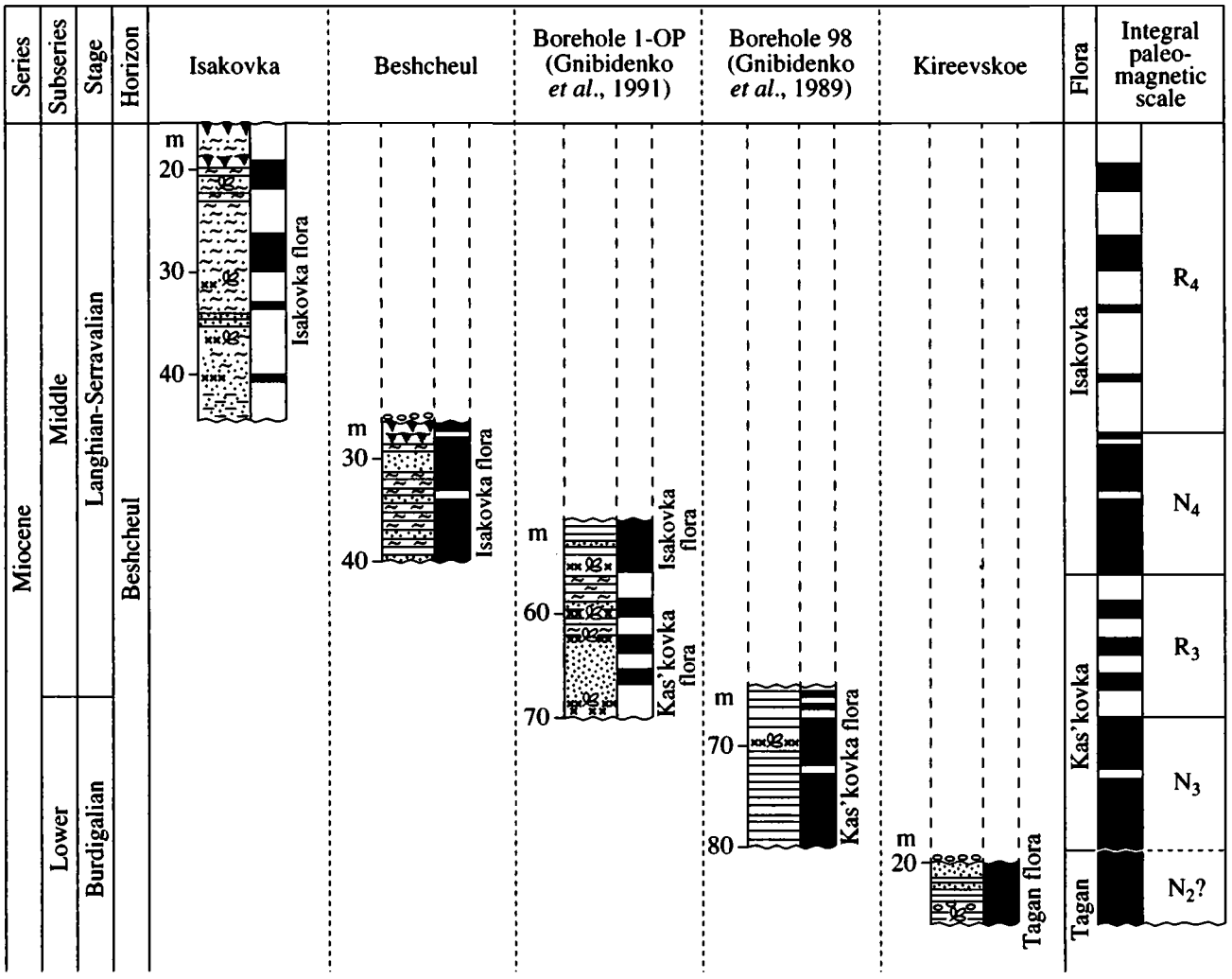

Fig. 3. The integral magnetostratigraphic section of the Beshcheul Horizon (symbols as in Fig. 2).

Kas'kovka time by thin forests locally grading into forest-steppes.

Paleomagnetic studies of layers with the Kas'kovka flora, which were recovered by some boreholes south of Barabinsk (Gnibidenko *et al.*, 1989) and near Omsk (Gnibidenko *et al.*, 1991), showed that they correspond to two magnetic zones of the normal (N) and reversed (R) polarities in the lower and upper parts of the section, respectively. Narrow intervals of the opposite polarity are detected in both zones (Fig. 3).

The Kireevskoe Settlement. The Tom'-Ob' region represents the eastern flank of the West Siberian plain. In this region, the younger Neogene layers were distinguished long ago near the Voronovo, Kireevskoe, and Zaobskii settlements at the Ob' riverside. In the Ob'-Tom' interfluve area, boreholes penetrated the Bolotnoe Formation overlain by the strata exposed near the Kireevskoe Settlement (the right-hand Ob' riverside) and termed as the Tagan Formation by M.P. Nagorskii in 1962. The latter is represented by the alluvial sequence composed of alternating inequigranular sand displaying diagonal bedding and bearing coal detritus and fragments, conglomerate, and sandy clay with limonite and siderite (Mizerov *et al.*, 1971). The clayey interbeds are typical of the middle and upper parts of the section; in one of them located in the upper part, S.G. Zhilin discovered abundant leaf impressions. The Tagan Formation rests upon the bluish gray, locally brownish clayey member (the apparent thickness 4–6 m) exposed near the water line or, more frequently, concealed under landslides and talus. The member corresponds to the known Kireevskoe Beds (Formation) of the Abrosimovka Horizon, which yielded leaf impressions and carpological and palynological assemblages.

The seed assemblages of the Tagan floral level are as yet little known. In general, the Tagan flora differs only slightly from those occurring in the underlying Abrosimovka Horizon. Forests of that time, however, lost many thermophilic elements, which were replaced by plant communities with *Alnus, Betula, Carpinus, Cory-*

lus, and other representatives of the temperate flora; they gradually migrated from watershed areas down to river valleys. The Tagan flora is characterized by the low abundance and diversity of Taxodiaceae, and by the absence of *Sparganium sibiricum* Dorof., *Sagisma turgida* P. Nikit., *Tavdenia, Evodia*, and other forms typical of the younger levels of the "Abrosimovka" flora distribution. Simultaneously, it is marked by the first occurrence of *Caulina graminea* Tzvel., *Eleocharis praemaximowiczii* Dorof., *Chenopodium hybridum* L., *Palaeoeuryale sukaczevii* Dorof., *Bunias*, and Asteraceae forms widespread in the southerly areas during the later periods. Based on the overall composition and main biogeographic parameters (in average, 44% of genera occur in modern communities, 43% are exotic, and 13%, extinct), the Tagan flora can be included, with the same degree of confidence, both into the Abrosimovka and younger Beshcheul floras. The autonomous Tagan floral level was identified (Nikitin, 1984) mainly on the basis of general geological history of the region: the Tagan Formation overlies the Kireevskoe Formation with a distinct erosional unconformity and probably corresponds to the initial phase of a new sedimentation cycle. Thus, the carpological assemblages of the Tagan floral level originate probably from the lowermost layers of the Beshcheul regional horizon, whereas assemblages of the Tagan and Kireevskoe levels, close in composition and age, represent, in our opinion, the final evolutionary stage of the Turgai flora in West Siberia. Moreover, it is not inconceivable that it is precisely this stage that corresponds to the middle Miocene climatic optimum that is well manifested in Europe and Pacific region, although the paleobotanical data cannot be regarded as direct indicators.

Spore-pollen spectra of the Tagan Formation are also transitional in composition (Merkulova, 1966b; Goloverov and Kabanova, 1984). On the one hand, they still contain relatively abundant pollen of broad-leaved arboreal plants (mainly the *Quercus* species close to the modern ones, *Ulmus* and Juglandaceae) associated with subordinate herbaceous and fruticulose taxa, and, on the other, pollen of broad-leaved plants is represented by a limited number of highly tolerant species, such as abundant *Betula* and, particularly, *Alnus* (sometimes, up to 45%). The group of coniferous is dominated by *Pinus* and *Picea* pollen, whereas Taxodiaceae pollen occurs sporadically in lower layers of the formation, and herbaceous pollen is sufficiently diverse (Poaceae, Chenopodiaceae, Ranunculaceae, Polygonaceae, Apiaceae, and others).

The section of the Tagan Formation near the Kireevskoe Settlement was sampled for the paleomagnetic study only in its 6-m-thick upper part, including the layer with leaf impressions collected but not yet determined. The studied interval corresponds to the magnetic N zone (Fig. 2).

DISCUSSION OF PALEOBOTANICAL DATA

The analysis of available paleobotanical materials suggests that deposits of the Beshcheul regional horizon accumulated during the epoch of significant changes in flora and vegetation in the southern part of the West Siberian plain, when the moderately thermophilic (mesophilic) Turgai forests of the late Oligocene–early Miocene were mainly replaced, under gradually deteriorating climatic conditions, by the forest-steppe landscapes. The map (Fig. 1) shows 154 localities of 222 seed assemblages corresponding to different floral levels (local zones) of the Beshcheul Horizon. As is seen, these assemblages cover virtually the entire study area of the Ob'-Irtysh interfluve. As stated above, the evolutionary characteristics of the assemblages allow three successive floral levels to be outlined in the Beshcheul Horizon. The lowest of them corresponds to beds bearing the Tagan flora. The seed assemblages of the Tagan level are well known only from several sections in the Tom'-Ob' region and interpreted as representing the final stage of the Turgai flora evolution in West Siberia. The next, well-represented Kas'kovka level corresponds to the initial stage in development of the transitional flora named as post-Turgai flora. Assemblages of the latter are widespread in southern areas of the plain. The relevant carpological assemblages are detected in many boreholes drilled in the eastern Barabinsk plain and Barnaul-Ob' region; recently, they were also discovered in some sections of the Omsk-Tara region adjacent to the Irtysh River. Whereas the share of modern endemic species in fossil seed assemblages from the uppermost layers of the Abrosimovka Horizon does not exceed 41–43%, in the Kas'kovka flora, their abundance is about 50% on average. This flora is already close to the known carpological assemblages from the Beshcheul Formation of the right-hand Irtysh riverside, although the latter are slightly less rich in Turgai relics. Simultaneously, the Kas'kovka level is marked by appearance of several species typical of the post-Turgai or even Recent floral stage. In the climate of the Kas'kovka time, there was a distinct tendency toward colder and more arid conditions; winters of that time were relatively longer, with the temperature probably about –10 to –12°C.

The third, Isakovka floral level is youngest one in the Beshcheul Horizon. The thorough analysis of fossil seed assemblages reveals that the flora from the Beshcheul Formation sections of the Irtysh River bank and adjacent areas of the Barabinsk and Ishim steppes differs from the Kas'kovka flora not only in provincial and taphonomic properties, but in terms of age as well. This flora is more advanced (in average, 55% of modern endemic genera) and, consequently, younger. It should be emphasized, however, that the Beshcheul Formation of the Omsk-Tara Irtysh region encloses, along with the flora of the Isakovka level, also the Kas'kovka flora. This can be exemplified, for instance, by the reference section of Borehole 1-OP drilled in southern suburbs of

Omsk, where the Beshcheul Formation hosts floras of both mentioned levels (Gnibidenko *et al.*, 1991).

DISCUSSION OF MAGNITOSTRATIGRAPHIC DATA

The paleomagnetic method not only enables subdivision and interregional correlation of intricate sections, but also allows the stratigraphic units to be dated and calibrated in accordance with the standard magnetogeochronological scale (Berggren, *et al.*, 1985, 1995). Recently, paleomagnetic characteristics in their continuous succession became known for different Upper Cenozoic stratigraphic subdivisions in some sections of West Siberia (Ob' and Irtysh river basins and their interfluve area), including the Neogene regional horizons, in general, and the Beshcheul Horizon, in particular. For instance, according to Gnibidenko *et al.* (1989), the Beshcheul Horizon of the Ob'-Irtysh interfluve area is represented by the Kas'kovka Beds, which correspond to two magnetic zones (from the base upward) with normal (N) and reversed (R) polarities; the zones are 10 and 6 m thick, respectively (Fig. 3). The comprehensive study of core section recovered by Borehole 1-OP in the Omsk-Irtysh region showed that the Beshcheul Horizon, corresponding here to beds with the Kas'kovka and Isakovka floras, also includes two zones displaying the reversed (R) and normal (N) magnetization of rocks (Fig. 3); in this case, they are 16 and 6 m thick, respectively (Gnibidenko *et al.*, 1991).

Magnetostratigraphic studies of the stratotype and reference sections of the Tagan and Beshcheul formations in the Ob' and Irtysh rivers basins provided first paleomagnetic data for the Tagan Formation (exposure near the Kireevskoe Settlement) and elucidated the stratigraphic range of beds with the Isakovka flora, thus, allowing us to correct the regional Neogene magnetostratigraphic scale of West Siberia. According to the former viewpoint, these beds corresponded to a zone of normal magnetization, whereas the recent paleomagnetic data divided this unit into two zones of normal (N) and reversed (R) polarity.

Paleomagnetic studies of all Beshcheul Horizon sections were conducted using oriented samples (cubes with facets of 2.4 and 2 cm) and following the standard procedure. Paleomagnetic sections were compiled according to directions of characteristic magnetization vectors determined using the component analysis. The latter was performed using the vector diagrams based on the results of the thermal stepwise demagnetization of samples.

Correlation of paleomagnetic sections characterizing beds with the Tagan, Kas'kovka, and Isakovka floras in the reference borehole sections and separate exposures was used as a basis for the regional magnetostratigraphic scale of the Beshcheul Horizon (s. l.), which includes five major magnetic zones of the normal and reversed polarity (Fig. 3). The position and succession of magnetic zones in the regional section were controlled by paleobotanical data, and their intervals were correlated with local biostratigraphic zones, or floral levels. Analysis of obtained magnetostratigraphic data shows that the revealed magnetic zones are of the regional type. With due account for zones earlier identified in the Abrosimovka Horizon (Gnibidenko *et al.*, 1996; Martynov *et al.*, 1997), we designated them from the base upward as N_2?, N_3, R_3, N_4, and R_4 magnetic zones (Fig. 3). In the course of time, these magnetic zones can be used under local names to compile the local schemes for other areas and to unite them subsequently into a general magnetostratigraphic scale of the whole region without changes in the numbering system, if an additional magnetic zone will be detected.

At the modern stage of paleomagnetic and biostratigraphic investigations, it is difficult to reliably correlate the succession of magnetic zones recognized in the Beshcheul Horizon of West Siberia with the standard magnetochronological scale. In our previous works (Gnibidenko *et al.*, 1996; Martynov *et al.*, 1997), it was shown that the accumulation of the Abrosimovka Formation terminated at the beginning of the Burdigalian Age. Taking into consideration the fact that the regional section of the Beshcheul Horizon begins with the Tagan Formation (N_2? magnetic zone) correlative with the Kushuk Beds, which enclose the Burdigalian mammal remains (*Homphotherium*) and are characterized by the normal polarity (our unpublished data), it can be assumed that the lower strata of the horizon accumulated as early as in the Burdigalian. The higher strata of the Beshcheul Horizon (Kas'kovka and Isakovka beds; N_3, R_3, N_4, and R_4 magnetic zones) continue the succession and probably correspond to the terminal Burdigalian, Langhian, and, partly, to the Serravalian. It should be emphasized that the orthostratigraphic fossil groups are missing from the Neogene deposits in West Siberia. Consequently, the confident application of the standard geochronological scale is hardly possible here, and the correlation discussed above should thus be considered as being preliminary.

CONCLUSION

The Beshcheul Horizon represents the boundary stratigraphic unit of the study region, which recorded significant environmental changes in West Siberia during the mid-Miocene time. The respective events included the activation of erosion induced by uplifts in mountains framing the plain, the development of the river system, the accumulation of a lateral succession of several alluvial formations, and the climate-related landscape reorganization. The latter affected particularly the development of flora and vegetation, which entered the post-Turgai evolutionary stage. The thorough stratigraphic study of the horizon using paleobotanical and magnetostratigraphic methods refined its time range: the horizon formation commenced probably in the Burdigalian Age approximately 20–19 Ma

ago and continued through the Langhian and Serravalian.

Provided that it is difficult or even impossible to select a universal stratotype for the given regional horizons and many formations, we have to use the composite reference sections integrating the laterally isolated stratigraphic units (Martynov, 1984). This is exactly the case with the stratotype proposed for the Beshcheul Horizon and formation of the same name: it appeared that the stratotype section located near the village of Isakovka probably represents only the upper part of the Beshcheul Formation, or the Isakovka Beds of the latter displaying the reversed polarity (R_4 magnetic zone). As evident from the paleomagnetic data, the section of the formation should be continued upward with beds containing the same Isakovka flora, but characterized by a normal polarity (N_4 magnetic zone). However, even in this case, the integral section of the horizon is still incomplete: the underlying layers with the Kas'kovka flora also correspond to two zones of the different polarity (R_3 and N_3 magnetic zones). Finally, beds with the Tagan flora can crown the stratigraphic section, if the normally magnetized deposits (N_3 and N_2? magnetic zones) enclosing the Kas'kovka and Tagan floras are not considered as one stratigraphic unit.

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Campanian-Maastrichtian Deposits in the Frontal Part of the Olyutor Zone (Southern Koryak Upland)

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Abstract—New dating of volcanogenic-siliceous oceanic (marginal-sea) deposits of the frontal Olyutor zone correlative with rocks of the Vatyna Group indicates that, in addition to Albian-Campanian, they include the younger Campanian-Maastrichtian horizons. The data allow us to contend that the oceanic and/or marginal-sea (Vatyna) and island-arc (Achaivayam) deposits, also dating back to the Campanian-Maastrichtian, were in lateral connections at that time. On the other hand, the analysis of new and published data shows that the Campanian-Maastrichtian island-arc deposits could probably overlap the pre-Campanian oceanic and/or marginal-sea sediments without any apparent disconformity.

INTRODUCTION

The frontal part of the Olyutor zone consists of the Cretaceous volcanogenic-siliceous-terrigenous complexes. Two principal stratigraphic units—the Vatyna Group and Achaivayam Formation—were recognized here by the late 1980s (Lipman, 1959; Zhamoida, 1972; Kazintsova, 1979; Alekseev, 1979). The study of volcanogenic rocks revealed two different genetic types of the Cretaceous deposits: those of oceanic and island-arc series (Bogdanov *et al.*, 1982; *Geologiya yuga...*, 1987; Chekhovich, 1993). In general, the Vatyna Group deposits were considered as similar to oceanic, whereas the Achaivayam Formation was classed with the island-arc one (Alekseev, 1979).

By the late 1980s, deposits of the Vatyna Group have been dated back to the Albian-Campanian, and island-arc deposits of the Achaivayam Formation to the Maastrichtian-Paleocene (Vishenvskaya, 1985; *Geologiya yuga...*, 1987). Data obtained recently indicate that, in addition to the Albian-Campanian, the Vatyna Group comprises younger deposits of the Campanian-Maastrichtian age (Palechek, 1997; Soloviev *et al.*, 1998), whereas the Achaivayam deposits likely accumulated in the Campanian (Palechek, 1997). This suggests some synchronism in the accumulation of oceanic (marginal-sea) and island-arc deposits, at least in the Campanian-Maastrichtian, as well as their lateral connections. Some researchers assume vertical relationships between units and assume that younger deposits of the Achaivayam Formation overlie with the older deposits of the Vatyna Group without an apparent disconformity (Astrakhantsev *et al.*, 1987; Kazimirov *et al.*, 1987; Shapiro, 1995). Our work aims to resolve this contradiction on the basis of new radiolarian dating of the Cretaceous deposits from the frontal part of the Olyutor zone. While simultaneously analyzing the published data on the age of Cretaceous deposits in the Olyutor zone, we took into account the available dating results for siliceous deposits associated with volcanic rocks of a known geochemical affinity, which allowed us to conceive this or that geodynamic environments of sedimentation.

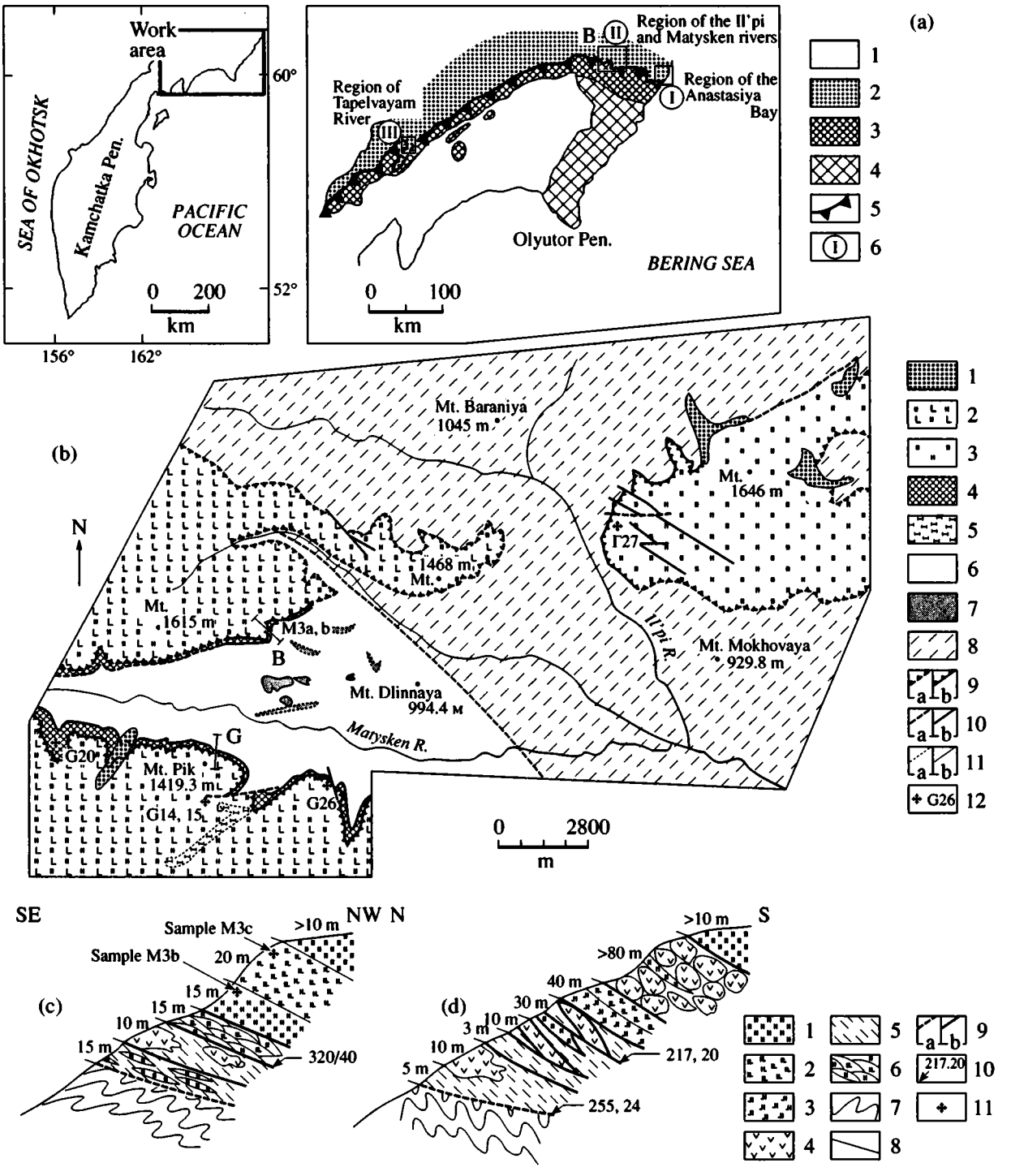
STRUCTURAL POSITION, COMPOSTION, AND AGE OF DEPOSITS IN THE FRONTAL PART OF THE OLYUTOR ZONE

The frontal part of the Olyutor zone extends along its northwestern boundary and is characterized by the imbricated nappe structure and a wide development of the Cretaceous deposits (Bogdanov *et al.*, 1982; Chekhovich, 1993) (Fig. 1a). The structure of the frontal part is defined by the Vatyna-Vyvenka regional thrust fault (Mitrofanov, 1977), more than 500 km long. Structural units strike here conformably with the Vatyna-Vyvenka thrust fault, along which the Olyutor zone deposits are thrust over the deposits of the Ukelayat trough.

The Cretaceous strata make up major subhorizontal slices (Chekhovich, 1993). We studied three regions within the frontal part of the Olyutor zone (Fig. 1a): the area of the Anastasiya Bay (the northern region I), the area of the upper courses of Il'pi and Matysken rivers (the northwestern region II) (Fig. 1b), and the area of the Tapelvayam River basin (the southwestern region III).

The northern region of the Anastasiya Bay. Allochthonous *volcanogenic-siliceous* and *volcanosedimentary* rocks of the northern Olyutor zone occur here in tectonostratigraphic successions consisting of imbricated tectonic slices of different ages (Palechek, 1997; Soloviev *et al.*, 1998).

Volcanogenic-siliceous deposits are exposed at the base of the allochthonous complex being represented



by basaltic pillow lavas, hyaloclastites, jasper with *Inoceramus* fragments, cherts, and aleuropelites. The late Turonian-early Campanian, Coniacian-middle Campanian, and Campanian-Maastrichtian radiolarian assemblages are macerated from siliceous rocks of various slices (Palechek, 1997; Soloviev *et al.*, 1998). Basalts of the volcanogenic-siliceous sequence are medium differentiated rocks (MgO/FeO_{tot} = 0.34–0.76). Judging from the content of TiO_2 (0.83–2.03%), Zr (76–160 ppm), and Y (30–47 ppm), basalts are similar to tholeiites of mid-oceanic ridges (MORB) and marginal seas (Hawkins, 1976; Sun *et al.*, 1979). Basalts show the moderate fractionation of LREE relative to HREE [$(La/Yb)_N$ ratios vary from 2.8 to 3.7] that, along with the behavior of major and other trace elements, allow us to class them with the E-type basalts of marginal sea basins (*Petrologiya i geokhimiya...*, 1987).

Fig. 1a. Distribution of Cretaceous deposits in the Olyutor zone (the study area is shown in the inset map: (1) Cenozoic deposits; (2) Cretaceous-Paleogene deposits of the Ukelayat zone; (3–4) the Cretaceous deposits of the frontal Olyutor zone (3) and the of the Olyutor Range region (4); (5) the Vatyna-Vyvenka thrust fault; (6) studied exposures

Fig 1b. Regional geologic scheme of the of Il'pi and Matysken rivers (compiled in collaboration with M.T. Brandon and G.I. Garver using materials by A.V. Ditmar, K.S. Ageev, A.S. Finogentov, and E.S. Alekseev): (1) Quaternary unconsolidated deposits; (2) aphyric pillow lavas of basaltic composition, cherts, and jaspers; (3) cherts and jaspers; (4) the Vatyna-Vyvenka thrust-fault zone; (5) blocks of dunites, wehrlites, and clinopyroxenites; (6) the tectonic melange zone; (7) basaltic blocks; (8) the Ukelayat flyschoid deposits; (9) thrust faults proved (a) and deciphered in aerial photographs (b); (10) steeply dipping faults proved (a) and hypothetical(b); (11) contacts proved (a) and hypothetical (b); (12) localities and numbers of samples with microfauna.

Figs. 1c–d. Fragments of tectonostratigraphic successions exposed at the left (C) and right (D) slopes of Matysken River: (1) green cherts; (2) black cherts; (3) red-brown jasper; (4) basaltic pillow lavas; (5) cataclastic aleuropelites; (6) siliceous rocks with boudinage; (7) the Ukelayat flyschoid deposits; (8) stratigraphic contacts; (9) tectonic contacts proved (a) and hypothetical (b); (10) bed attitude; (11) localities and number of samples with microfauna.

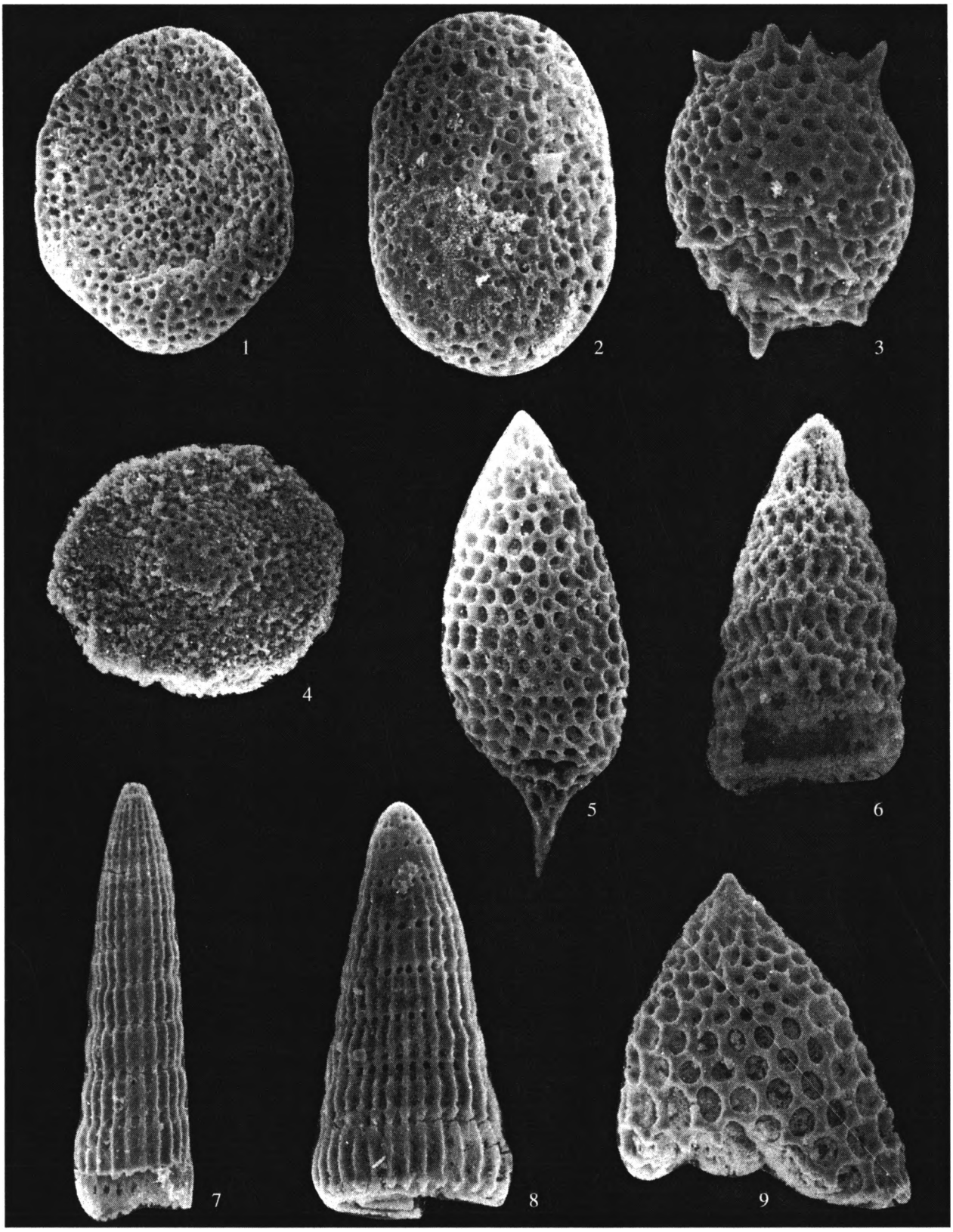
Volcanosedimentary deposits are widespread within the area of Anastasiya Bay. According to our results, basalts are subdivided, with a fair degree of convention, into three sequences: (1) *volcanogenic (lower)*; (2) *volcanogenic-terrigenous (middle or "transitional")*; and (3) *siliceous-terrigenous (upper)*. The lower volcanogenic sequence includes clinopyroxene-phyric amygdaloidal basalts, lava breccia of basaltic composition; lava clastics in a graywacke matrix and lenses of basaltic andesites are rare. The volcanosedimentary sequence overlaps with tectonic contact the volcanogenic-siliceous deposits described above. *The volcanogenic-terrigenous* sequence ("transitional", middle) is characteristic of isolated allochthonous slices. It includes graywacke sandstone beds, monomictic basaltic breccias, basalts, lava breccias, and siliceous rocks with the admixture of terrigenous material. *The siliceous-terrigenous sequence* (upper) is represented by graywacke sandstones, aleuropelites, cherty siltstones, green and black cherts. The sequence overlies deposits of the volcanogenic sequence. Radiolarian assemblages from cherts of the siliceous-terrigenous sequence indicate that volcanosedimentary deposits of the Anastasiya Bay area comprise the Campanian-Maastrichtian horizons (Palechek, 1997; Soloviev *et al.*, 1998). Basalts and lava breccias from volcanosedimentary deposits vary considerably in composition: TiO_2 (0.56–1.16%), Zr (8–82 ppm), Y (4–32 ppm), and MgO/FeO_{tot} = 1.03–1.64. By geochemical parameters, these occupy an intermediate position between typical tholeiitic and calc-alkali series. Basalts exhibit a higher fractionation of LREE relative to HREE [$(La/Yb)_N$ = 3.2–7.7] that is typical of the island-arc rocks. By geochemical characteristics, the basalts are correlative with those calc-alkali series moderately rich in potassium (*Petrologiya i geokhimiya...*, 1987; Frolova *et al.*, 1989).

The northwestern region (upper courses of the Il'pi and Matysken rivers). Several structural elements are distinguished in the upper courses of the Il'pi and Matysken rivers (Fig. 1b; (from the base upward): (1) *autochthon* of flyschoid deposits of the Ukelayat trough; (2) the tectonic melange zone; (3) the Vatyna-Vyvenka thrust fault; and (4) *allochthon* of volcanogenic-siliceous deposits of the Olyutor zone. A similar structure is characteristic of the northern part of the Olyutor zone (Astrakhantsev *et al.*, 1987).

The tectonic melange zone is chaotic in structure. The primary nature of the sequence is undeterminable here, because the rocks are subjected to intense tectonization; it likely was an olistostrome originally. The matrix is composed of black aleuropelites with rare and thin interlayers of fine-grained sandstone. Blocks are represented by phyric and aphyric basalts, lava breccias, and basaltic hyaloclastites, all rarely intercalated with black and green cherts. In the northern slope of Matysken river valley, there are exposed blocks of magmatic rocks: porphyroid melanocratic gabbro, leucocratic gabbro-diorite, and rare ankaramite-like gabbro. Chilled contacts of basaltic flow crosscutting the terrigenous flyschoid rocks of the Ukelayat trough are described at the foot of Mt. Dlinnaya. Near the contact zone, siltstones are transformed into black hornfels, and magnophyric basalts become microphyric to aphyric in the quenching zone. The sequence is of an imbricated structure. Further south, a similar rock sequence of analogous structure was regarded as an olistostrome, whose age was determined on the basis of such benthonic foraminifers as Maastrichtian (Mitrofanov and Sheludchenko, 1981).

The Vatyna-Vyvenka thrust fault zone (from 50 to 150 m thick) is characterized by imbricated structure. Tectonic slices of the zone are composed of rocks of the Olyutor zone and of deposits of the Ukelayat trough. The zone is especially distinct in slopes of the Matysken River valley (Fig. 1b). The slices bordered by subhorizontal thrust-fault planes consist of black and green siliceous rocks displaying boudinage, aphyric basalts, lava breccias, black cataclastic aleuropelites, and more rare sandstone beds. The slices vary in size from a few to dozens of meters.

Volcanogenic-siliceous deposits occupying the highest hypsometric level and often constituting the isolated tectonic slices are represented by basaltic pillow lavas, lava breccias, basaltic hyaloclastites, and siliceous rocks. In the studied region, the latter are diverse and represented by red jasper beds enclosing interlayers with broken *Inoceramus* shells and intercalated with massive black, green and gray cherts. Volcanogenic-siliceous rocks are highly disintegrated, and it is



Radiolarians from volcanogenic-siliceous deposits in upper courses of the Il'pi and Matysken rivers: (1) *Phaseliforma carinata* Pessagno, ×300, Sample G26; (2) *Phaseliforma laxa* Pessagno, ×200, Sample G26; (3) *Lithomespilus mendosa* (Krasheninnikov), ×300, Sample G26; (4) *Pseudoaulophacus* cf. *lenticulatus* (White), ×200, Sample M3/v; (5) *Stichomitra livermorensis* (Campbell & Clark), ×250, Sample G26; (6) *Wildeus punctulatus* (Pessagno), ×300, Sample G14; (7) *Dictyomitra andersoni* (Campbell & Clark), ×200, Sample G26; (8) *Dictyomitra densicostata* Pessagno, ×300, Sample G26; (9) *Clathrocyclas* cf. *hyronia* Foreman, ×200, Sample G26.

Radiolarians from the frontal part of the Olyutor zone

Species	Region of the Anastasiya Bay (Soloviev *et al.*, 1998)			Upper courses of the Il'pi and Matysken rivers			Region of the Tapelvayam River		
	Sample number								
	4	29	A12	M3/v	G26	G27	3T	17T	18T
Phaseliforma carinata Pessagno	*	*	*	*	*				
P. subcarinata Pessagno		*							
P. cf. *laxa* Pessagno					*				
P. cf. *meganosensis* Pessagno		*							
Phaseliforma sp.						*	*		
Orbiculiforma sp.	*			*			*	*	*
Praestylosphaera pusilla (Camp. & Clark)	*	*	*						
P. hastata (Camp. & Clark)		*	*	*					
Praestylosphaera sp.						*			
Lithomespilus mendosa (Krasheninnikov)			*	*	*	*			
Pseudoaulophacus cf. *lenticulatus* (White)							*		*
Pseudoaulophacus sp.				*				*	
Alievium sp.	*								
Prunobrachium sp.									*
Spongurus sp.				*					
Archaeospongoprunum sp.	*							*	
Actinomma sp.	*								
Stichomitra livermorensis (Camp. & Clark)	*	*	*		*		*		
S. cf. *shirshovica* Vishnevskaya		*							
Stichomitra sp.						*			*
Cornutella californa Camp. & Clark	*		*				*		*
Amphipyndax stocki (Camp. & Clark)	*	*	*	*	*		*	*	*
A. stocki var. B Vishnevskaya		*							
A. streckta Empson–Morin	*	*	*						
Theocampe? sp.							*	*	
Bathropyramis sp.	*								
Archaeodictyomitra regina (Camp. & Cl.)		*			*				
Dictyomitra andersoni (Camp. & Cl.)					*				
D. densicostata Pessagno	*	*	*		*	*			
D. multicostata Zittel	*		*	*	*				
Dictyomitra sp.								*	
Clathrocyclas hyronia Foreman	*	*	*	*	*				
C. tintinnaeformis Camp. & Clark		*							
Clathrocyclas sp.						*			
Xitus cf. *asymbatos* (Foreman)	*								
Xitus sp.			*						

* Species presence in samples.

impossible to restore the original sequence of their thin and isolated slices (Figs. 1b, c, d). Different horizons of volcanogenic-siliceous deposits occur over the thrust fault zone, for instance, basaltic lavas with chert lenses in the southern and northern slopes of the Matysken River valley and siliceous rock further to the north and east (Fig. 1b).

Radiolarians of different preservation degree were extracted from siliceous rocks of the volcanogenic-sili-

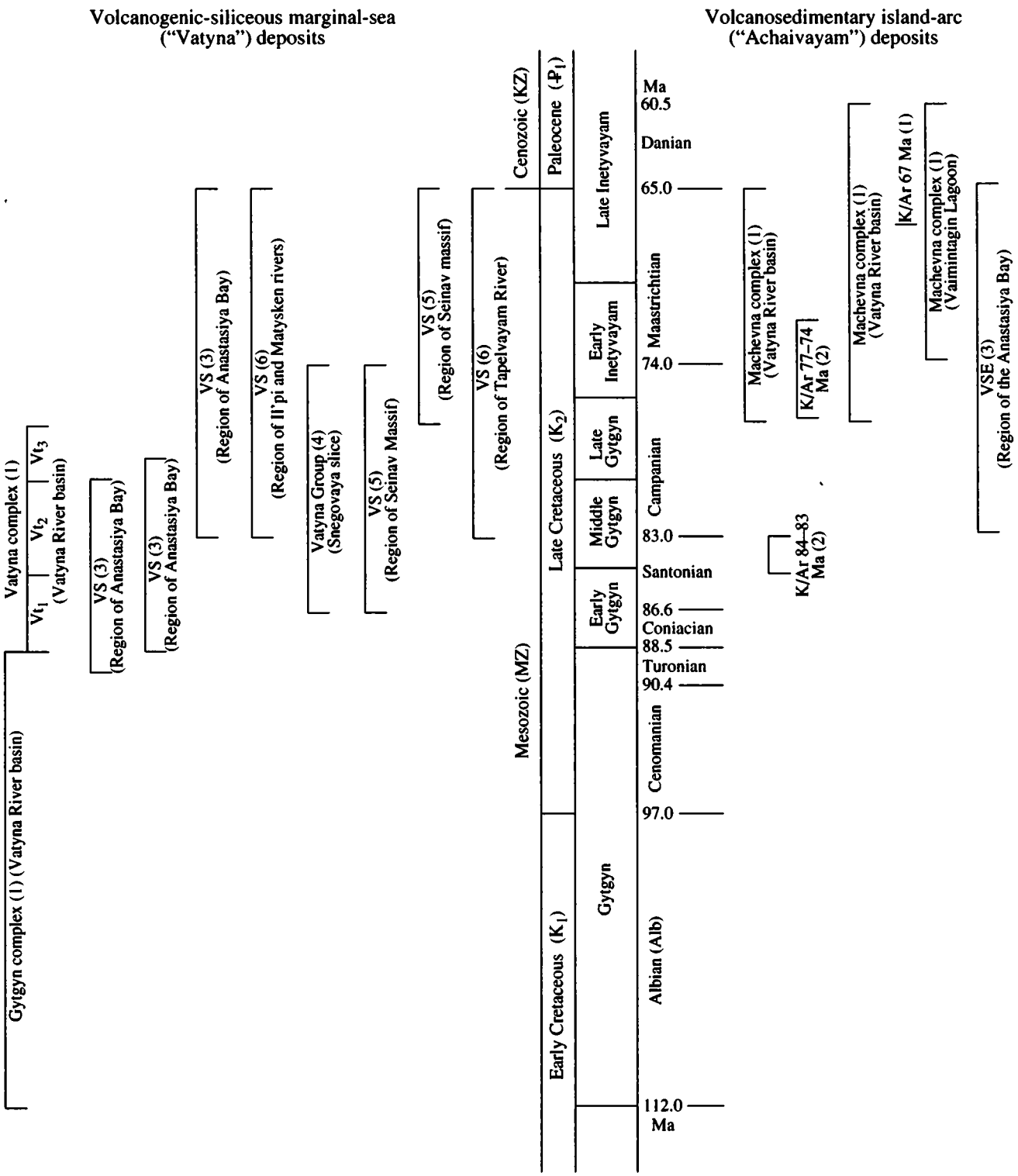


Fig. 2. Correlation scheme of volcanogenic-siliceous marginal-sea ("Vatyna") and volcanosedimentary island-arc ("Achaivayam") deposits. The geochronological scale is calibrated on the basis of radiolarian assemblages identified the Olyutor zone (Vishnevskaya, 1985). Shown in the scheme are time ranges of tectonic slices, where siliceous deposits associate with volcanic rocks of the marginal-sea (on the left) or island-arc (on the right) types: (VS) volcanogenic-siliceous and (VSE) volcanosedimentary sequences (Palechek, 1997, and this work). K/Ar dates are shown for basaltic rocks of the Vatyna River basin. Numbers in parentheses indicate data sources: (1) *Geologiya yuga...*, 1987; (2) Alekseev, 1979; (3) Soloviev *et al.*, 1998; (4) Astrakhantsev *et al.*, 1987; (5) Kravchenko-Berezhnoy *et al.*, 1993; (6) this work.

ceous slices and also from the Vatyna-Vyvenka thrust-fault zone. Radiolarian assemblages are of a low taxonomic diversity and abundance (see the plate and table). Eight radiolarian taxa have been identified in a very representative sample (G26) as being represented by a few dozen specimens. Their assemblage (G26) includes *Phaseliforma carinata* Pessagno, *Ph. laxa* Pessagno, *Lithomespilus mendosa* (Krasheninnikov), *Stichomitra livermorensis* (Campbell & Clark), *Archaeodictyomitra regina* (Campbell & Clark), *Dictyomitra andersoni* (Campbell & Clark), *D. densicostata* Pessagno, *D. multicostata* Zittel, and *Clathrocyclas* cf. *hyronia* Foreman (table and plate). The assemblage suggests that volcanogenic-siliceous deposits in the upper courses of the Il'pi and Matysken rivers comprise the late Campanian-Maastrichtian horizons. We should point to the distinct deformation of some radiolarian skeletons in the assemblage (G26) from the thrust-fault zone. This is most typical of *Phaseliforma carinata* Pessagno and less characteristic of nasselarians *Archaeodictyomitra regina* (Campbell & Clark). The radiolarian assemblage is similar in taxonomic composition, but less diverse than the coeval assemblage from the Anastasiya Bay region (Palechek, 1997; Soloviev *et al.*, 1998). Another radiolarian assemblage of the Campanian and, probably, early Maastrichtian age was identified earlier by Kazintsova (1979) in thin sections of siliceous rocks from the Il'pi River basin. In some features, it is less similar to our assemblage than the late Vatyna radiolarians (Vishnevskaya, 1985).

The volcanosedimentary deposits characteristic of the Anastasiya Bay region are not found in the region under discussion. These deposits have likely been eroded.

The southwestern region (the Tapelvayam River basin). A large tectonic outlier of the Olyutor zone deposits (Mitrofanov, 1977) is surrounded here by flyschoid deposits of the Late Cretaceous-Paleogene Ukelayat zone representing the autochthon. The paraautochthon is composed of the Maastrichtian "olistostrome" sequence (Mitrofanov and Sheludchenko, 1981). The outlier represents a packet of allochthonous slices of different deposits (Astrakhantsev *et al.*, 1991; Kravchenko-Berezhnoy *et al.*, 1993). The lower slice is composed of volcanogenic-siliceous rocks of the Vatyna Group of the Santonian-Campanian (Astrakhantsev *et al.*, 1991) or Campanian-Maastrichtian age (Kravchenko-Berezhnoy *et al.*, 1993). The middle slice is represented by tuffaceous-siliceous deposits of the Maastrichtian-Paleocene Achaivayam Formation. The upper slice is made up of ultramafic and gabbroid rocks of the Seinav, Galmoenan and Imlan massifs.

We studied *the volcanogenic-siliceous deposits* in the upper courses of the Oginravayam River, where they are thrust over the Ukelayat flysch. Their sequence includes basaltic pillow lavas, hyaloclastites, basic lava breccias, siliceous rocks, and cherty aleuropelites. The radiolarian assemblage of low taxonomic diversity and of a medium preservation degree was detected in siliceous rocks (see table). The assemblage consists of *Stichomitra* cf. *livermorensis* (Campbell & Clark), *Cornutella* cf. *californica* Campbell & Clark, *Ampipyndax stocki* (Campbell & Clark), and *Pseudoaulophacus* cf. *lenticulatus* (White). It suggests the Campanian-Maastrichtian age of enclosing deposits. Radiolarian assemblages similar in taxonomic composition were identified earlier by Vishnevskaya in siliceous deposits of the Seinav Mountain area in the southwestern region (Kravchenko-Berezhnoy *et al.*, 1993).

DISCUSSION

(1) In general, the Campanian-Maastrichtian radiolarian assemblages from volcanogenic-siliceous deposits of different areas of the frontal part of the Olyutor zone are similar in their taxonomic composition and correlate well with one another (table). The studied assemblages are comparable to the late Vatyna and early Inetyvayam radiolarian faunas (Vishnevskaya, 1985). They are also correlative with the radiolarian assemblages from the Upper Cretaceous deposits of California (Campbell and Clark, 1944; Pessagno, 1976), DSDP Site 275 (Pessagno, 1975), and New Zealand (Hollis, 1997).

(2) Volcanogenic-siliceous deposits in all studied areas of the frontal part of the Olyutor zone are similar in the composition and structural position. Along with older beds, they include the Campanian-Maastrichtian horizons, and this allows their correlation within the time interval in question. The rocks are also comparable in composition to oceanic and/or marginal-sea deposits of the "Vatyna Group" (Bogdanov *et al.*, 1982; *Geologiya yuga...*, 1987; Chekhovich, 1993), the formation of which could have lasted correspondingly until the latest Cretaceous. Volcanosedimentary deposits of the northern Anastasiya Bay region comprise the Campanian-Maastrichtian horizons as well (Palechek, 1997; Soloviev *et al.*, 1998), and they are correlative in composition with island-arc deposits of the "Achaivayam Formation" (Chekhovich, 1993) which thus accumulated their beginning from the Campanian.

(3) The correlation scheme for the Cretaceous oceanic (marginal-sea) and island-arc deposits of the Olyutor zone (Fig. 2) is compiled on the basis of new age determinations and published data. The scheme suggests some important consequences. Obviously, the Campanian-Maastrichtian island-arc deposits overlie, most likely without any apparent disconformity, the pre-Campanian oceanic and/or marginal-sea sequences, as it has been reported in a series of publications. Such a situation is likely when an island arc develops on the oceanic crust basement (Astrakhantsev *et al.*, 1987; Kazimirov *et al.*, 1987). On the other hand, simultaneous formation of island-arc and oceanic (marginal-sea) deposits during the Campanian-Maastrichtian (Fig. 2) allows us to suspect the primary lateral connections between them and to interpret their con-

tacts in the modern structure as a result of the later tectonic imbrication.

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Late Miocene Brackish-Water Mollusks and the History of the Aegean Basin

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Abstract—Brackish-water mollusks are studied from several sections in northern Greece (Macedonia, the Serres basin) and the Aegina Island. Six of the 22 species of bivalve mollusks and one subgenus of cardiids are identified as new taxa: *Cerastoderma macedonicum, Lymnocardium (Euxinicardium) prominens, L. (E.) serresense, L. (Hellenicardium) hellenicum* subgen. et sp. nov., *L. (H.) strimonense*, and *Pseudocatillus submedius.* Ten species and almost all genera and subgenera (except *Cerastoderma, L. (Hellenicardium)*, and *Mactra*) are identical to the Pontian taxa of the Eastern Paratethys. Three species, one subspecies and all genera and subgenera (except the new taxa) were reported from the Messinian of Italy and Spain. In the Serres basin, deposits of the underlying Dafni Formation enclose abundant assemblages of marine mollusks, while the overlying sediments contain the marine microfauna and oysters. According to mollusks, the brackish-water deposits of the section stand in direct correlation with Pontian strata the Paratethys, at least with a part of them, on the one hand, and probably corresponds to the upper Messinian of the Mediterranean, on the other. According to the paleogeographic data, the Aegean basin was not a gulf of the Paratethys, but either a part of the Messinian basin, or an intermediate basin, where the Pontian fauna was formed.

Key words: *stratigraphic events, upper Miocene, Messinian, Pontian, Greece, Aegean basin, bivalve mollusks*

INTRODUCTION

Specific brackish-water mollusks of the Pontian type have long been known in the Mediterranean area from sections studied in Greece, Italy, southern France, and Spain (Almera, 1889; Gillet, 1965; Gillet and Geissert, 1971; Papp *et al.*, 1978; data of Stevanovic from *Pontien...*, 1989; Roegl *et al.*, 1991; Geronimo *et al.*, 1991). However, researchers, who often studied the mollusks, either had no previous experience with the typical Pontian faunas, or considered the Pontian of the Paratethys only on the basis of unrepresentative occasional Mediterranean collections.

The present research is based on the mollusk collection (about 500 specimens) obtained by S.V. Popov in the course of two months' work in Greece, the project *Paleogeographic assemblages and correlation of the Tethys and Paratethys Neogene deposits*. His work was supported by the NATO Fellowship Program. In this work, Popov was responsible for description of the studied sections, their correlation, and determination of marine fauna; L.A. Nevesskaya described brackish-water mollusks with the participation of N.P. Paramonova (the description of *Mactra faugeresi*). Associated gastropod and foraminifer assemblages are described according to determinations by L.B. Il'ina and T.N. Pinchuk (Scientific Technical Center Kuban'gasprom), and nannoplankton was studied by L.A. Golovina (Geological Institute, Russian Academy of Sciences, Moscow).

RESULTS

In Greece (Fig. 1), the brackish-water mollusks of the Pontian type were found in four areas: the Souvalia site of the Aegina Island (Roegl *et al.*, 1991), exposures of the Tracones Formation south of Athens (Papp *et al.*, 1978), a series of localities in the Strimon River basin (Gramann and Kockel, 1969), and the region of Saloniki (Gillet and Gissert, 1971; Steffens *et al.*, 1979; data of Stevanovic from *Pontien*, 1989). Except for the fourth site, all of them were visited, observed, and sampled in detail.

(1) The Souvalia section of the Aegina Island is well exposed and consists of the relatively deep-sea marly facies. The macrofauna is represented here only by brackish-water mollusks, whose specific forms, such as *Paradacna abichi, Pseudocatillus pseudocatillus, Lymnocardium (Euxinocardium)* cf. *nobile*, and *Pontalmyra incerta* are also characteristic of the Pontian. This fauna was found at three levels (Fig. 2; Table), where it is rather scarce and poorly preserved. The composition of benthic foraminifers is more diverse here than in the Pontian of the Eastern Paratethys; in addition to *Ammonia beccarii*, their assemblage includes *Quinqueloculina* sp., *Guttulina* sp., *Elphidium* aff. *reginum, E. macellum, Discorbis* sp. (data by

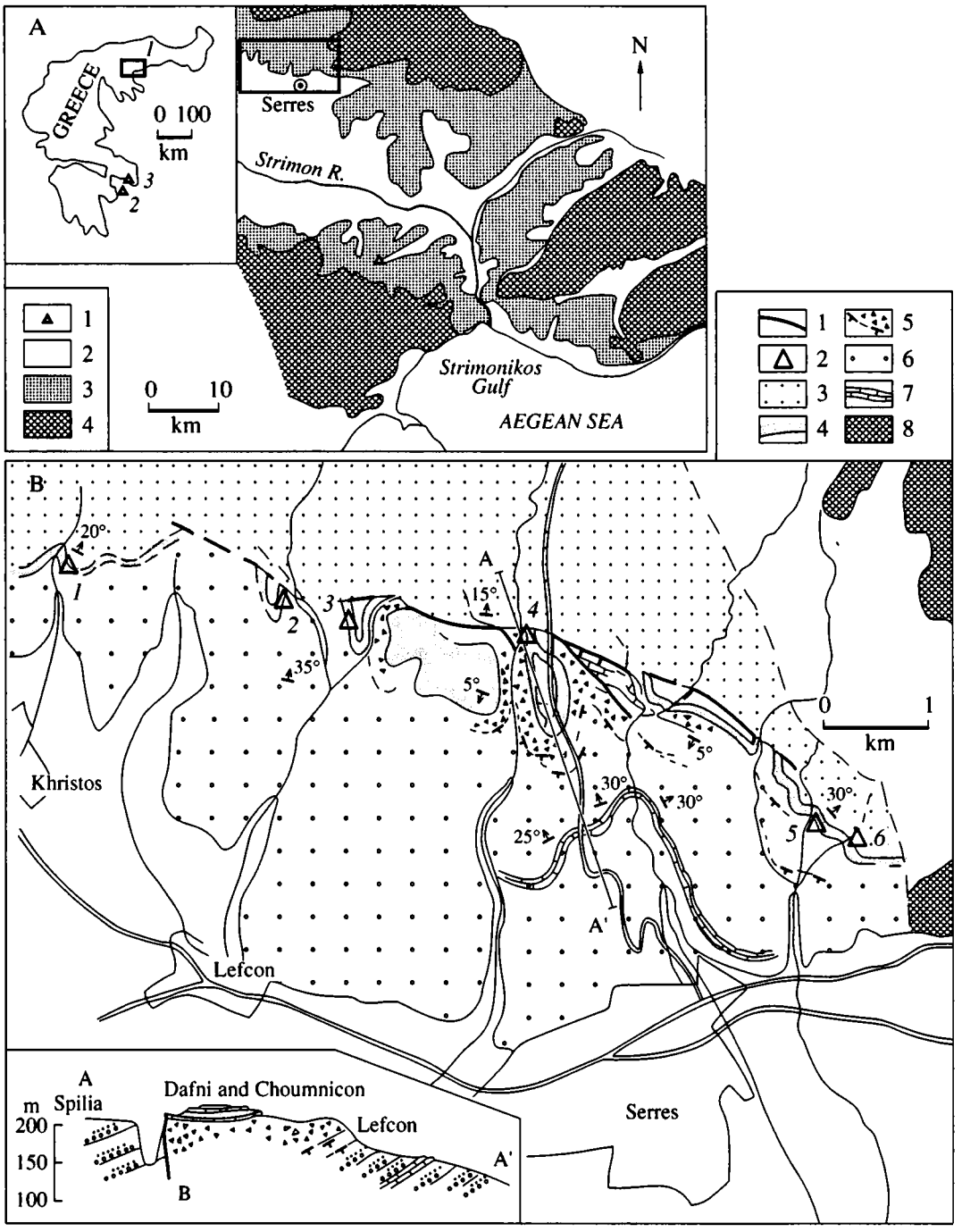


Fig. 1. Geological scheme of the Serres basin and sampling sites of the studied mollusks (compiled by S.V. Popov using the geological survey data). (A): (1) sampling site; (2) Quaternary, (3) Neogene, and (4) pre-Neogene deposits. (B): (1) faults; (2) sampling site; (3) Spilia Formation; (4) Dafni and Choumnicon formations; (5) breccia or red stone, (6) conglomerate, and (7) limestone of the Lefcon Group; (8) magmatic rocks.

T.N. Pinchuk from Popov *et al.*, 1996). The brackish-water sediments overlie continental deposits, where a *Hipparion* tooth was found (Roegl *et al.*, 1991). The overlying marine sediments bear foraminifers of genera *Polymorphina, Cassidulina, Anomalina, Globigerina, Globigerinoides, Bolivina*, and *Bulimina* (Popov *et al.*, 1996). According Roegl *et al.* (1991), this level also contains the nannoplankton assemblage with *Ceratolithus acutus*, pteropods, and also spicules and larvae of echinoderms.

(2) In the type locality of the Tracones Formation, fauna of brackish-water mollusks of the Pontian type

occurs in large rock blocks embedded in the olistostrome matrix. Some of the blocks contain only the brackish-water fauna, while others from the same stratigraphic level yield the euryhaline marine fauna of pectinids and colonial corals; and blocks of the third type bear the impoverished assemblages of semimarine fauna. Mollusks are represented by casts and molds. Their brackish-water assemblage contains the sharply keeled *Congeria* sp. (or *Dreissena* ex gr. *polymorpha*), *Pseudocatillus pseudocatillus, Mactra* cf. *faugeresi*, and terrestrial gastropods *Helix* sp.

(3) The deposits with the brackish-water fauna in the Strimon river basin are exposed on the northeastern, southern and eastern flanks of this synclinorium (Gramann and Kockel, 1969). The most abundant and well-preserved fauna was detected from the sections north of the town of Serres, where it is buried in sands, sandstones and conglomerates, whereas the section studied in the southern flank near the Oreskia Village is more calcareous in composition and contains mainly the casts and molds of molluscan shells.

The Serres Basin

In the basin, the Neogene deposits with marine and brackish-water fauna are exposed in the southern termination of the syncline located to the north of Serres (Fig. 1). Their beds stretch as a narrow band for 15 km and show mostly the monoclinal bedding disturbed by faults in the middle part.

The section begins with a thick sequence of conglomerate, lignite, and red beds of continental origin, which are distinguished as the Lefcon Group. In the group, there are known Lefcon and Meramena localities of vertebrate remains (*The Vertebrate...*, 1995). The age of the latter locality is determined as intermediate between the Turolian and Ruscinian. On the geographical scheme, it is shown in a tectonic window within the syncline core. The field observations, however, showed that this interpretation is insufficiently convincing, and it is not inconceivable that the finds of Meramena vertebrates originate from the overlying Spilia Formation. The lignite intercalations in the uppermost part of the Lefcon Group are uraniferous. The coarse detrital breccia at the roof is variably thick, locally up to 50–60 m.

Above the latter, the marine sediments (Georgeos Group) are represented by sand with a subordinate conglomerate, marl, and clay interbeds. The lower part of the section (Dafni Formation) is 30–35 m thick and contains the euryhaline marine fauna. The following species were recovered from basal beds: *Pecten benedictus, Arca* sp., *Anadara* sp., *Ostrea edulis, Crassatella* sp., *Acanthocardia* sp. (samples 1.1 and 1.3; Fig. 2). The Dafni upper conglomerate (samples 1.6 and 1.7) has also the euryhaline fauna: *Arca* cf. *noae, Modiolus adriaticus, Brachydontes* sp., *Crassostrea* sp., *Chama gryphoides, Cardita calyculata, Acanthocardia* sp., *Cerastoderma scabrum, Venus (Ventricoloidea)* cf. *multilamella, Timoclea ovata*, and *Corbula gibba*. However, between these beds, there are interlayers with mollusk assemblages impoverished to a different degree and containing, in addition to euryhaline marine forms, the brackish-water *Congeria* ex gr. *amygdaloides* and *C.* ex gr. *subcarinata*. It is interesting to note that the latter forms are characterized by a poor development of the apophysis, which may even disappear; in other words, they are transitional in morphology to *Dreissena*. Certain beds contain assemblages similar to those of the early Maeotian (Sample 1.5); other beds enclose the monotaxonic *Mactra* coquinas similar to those of the late Sarmatian (Khersonian; Sample 1.4). Assemblages from these interbeds

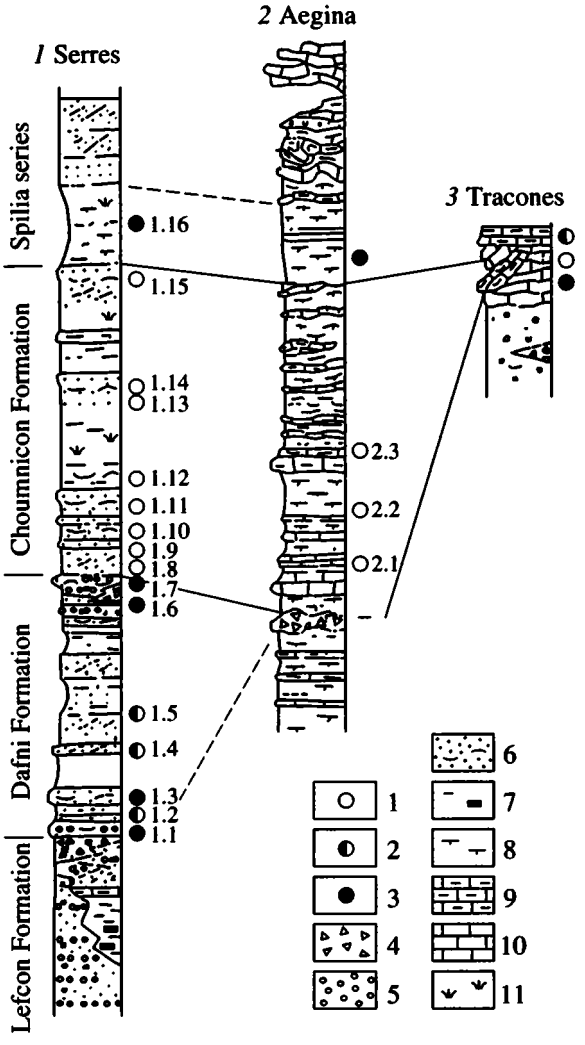


Fig. 2. Correlation scheme of the studied sections and sampling levels of fauna: (1) brackish-water mollusks; (2) impoverished assemblage; (3) euryhaline assemblage; (4) breccia; (5) conglomerate; (6) sand or sandstone with molluscan remains; (7) carbonate-free clay with lignite; (8) calcareous clay; (9) marl; (10) sandstones; (11) unexposed interval.

Geographic distribution of brackish-water molluscan species collected in the Serres basin and Aegina Island

Species \ Basins	Spain	Northern Italy	Sicily-I.	Greece			Pannonian basin	Dacian basin	Euxinian basin	Occurrence in the Aegean sections (see levels in Fig. 2)
				Aegina I	Trilophos	Serres				
Cerastoderma macedonicum Nev. sp. n.						+				1.9; 1.11
Lymnocardium (*Euxinicardium*) *nobile* Stef., 1896				cf	+	+		+	+	1.11; 1.12; 2.2; 2.3
L.(*E.*) *prominens* Nev., sp. n.						+				1.15
L.(*E.*) *serresense* Nev., sp. n.					cf	+				1.12
L.(*E.*) aff. *nobile* Stef., 1896						+				1.8; 1.12
L.(?*E.*) *almeri* Gillet, 1965	+					cf				1.9
L.(*Hellenicardium*) *hellenicum* Nev., sp. n.						+				1.13
L.(*H.*) *strimonense* Nev., sp. n.						+				1.11
Eupatorina littoralis (Eichw., 1850)			+		+	+		+	+	1.8; 1.11–1.14
Pontalmyra incerta chiae (Almera, 1894)	+	+		+	+	+				1.11–1.14; 2.2
Pseudocatillus pseudocatillus (Barb. de Marny, 1869)	+		+	+	+	+		+	+	1.8; 1.9; 1.12; 1.14; 2.2
P. submedius Nev., sp. n.						+				1.8
P. subdentatus (Deshayes, 1898)						+		+	+	1.10; 1.11; 1.14
P. multicostatus (Gillet, Geiss., 1971)				cf	+	+				1.8; 1.9; 1.11; 2.1
Paradacna abichi (R. Hoern., 1874)				+	+	+	+	+	+	1.8; 1.9; 1.12; 2.1–2.3
Abra tellinoides (Sinz., 1884)				cf	+	cf		+	+	1.8; 1.11–1.14
Parvivenus widhalmi (Sinz., 1877)				+	+	+		+	+	1.8; 1.9; 1.11–1.14; 2.1
Mactra faugeresi (Gill., Geiss, 1971)		?	cf	cf	+	+				1.9; 1.12; 1.13; 2.1; 2.3
Dreissena ex gr. *rostriformis*			+			+				1.8–1.11; 1.14
Dreissena simplex (Barb. de Marny, 1869)					+	+				1.8; 1.12
Congeria ex gr. *amygdaloides*					+	+				1.5; 1.8
Congeria subcarinata (Desh., 1838)					+	+				1.1; 1.6
Melanopsis ex gr. *impressa Krauss*					+	+				1.11; 1.13; 1.15

probably gave rise to the idea that subdivisions of the Eastern Paratethys might be distinguished in this area (Gramann and Kockel, 1969; Stevanovic and Il'ina, 1982; data of Stevanovic from *Pontien...*, 1989; Rust, 1997).

The overlying Choumnicon Formation is characterized by the brackish-water mollusk assemblage with numerous genera and species common with the Pontian taxa of the Eastern Paratethys: *Eupatorina littoralis*, two *Pseudocatillus* species, five of *Lymnocardium (Euxinicardium)* species, and *Paradacna abichi* (Table). Most of the species differ, however, from those of the Pontian and coexist with *Mactra* and *Cerastoderma* forms unknown in the Pontian. Six species and one *Lymnocardium* subgenus are described below as new. Two species, namely *Mactra faugeresi* and *Pseusocatillus multistriatus* are described for the first time from the Trilophos locality south of Saloniki (Gillet and Gessert, 1971), being also unknown in the Paratethys. Six-seven species and almost all of the detected genera are known as typical of the brackish-water Messinian deposits in the northern Mediterranean, Spain, northern Italy, Sicily, and Cyprus (Table).

Despite the abundant material, it is difficult, however, to positively state that changes in the brackish-water molluscan assemblages progress upward in the section, because many species are widespread throughout the section, and the distribution of rare forms may be accidental (Fig. 2; Table). We can only point out that the *Cerastoderma* and *Mactra*, which distinguish these assemblages at a generic level from the Pontian assemblages, are found only in the lowermost part of the Choumnicon Formation. Concurrently, *Mactra* forms reach the higher level in the Souvalia section. In the four samples from the Choumnicon Formation, L.A. Golovina identified two nannoplankton species *Braarudosphaera bigelowi* and *Perfocalcinella fusiformis*. Such a low diversity of nannoplankton indicates

stress environments for this group. The last species was known earlier only in the Paratethys from the lower Pannonian deposits of Hungary; later, it was found in the lower-middle Sarmatian deposits of Moldova and the Taman' Peninsula.

The clay member at the base of the overlying Spilia Formation (or at the roof of the Choumnicon Formation) encloses oyster shells and point to a short-term reinstallation of the marine environment (like in the Aegina Island section). A similar transition from the brackish-water sediments containing limnocardiids to marine sediments of the nannoplankton *Ceratolithus acutus* Zone was also observed in the sedimentary sequences of the Saloniki region (Steffens *et al.*, 1979). The Choumnicon Formation is about 40 m thick.

The upper Spilia Formation is composed of coarse-grain continental sands and small-pebbled conglomerates with the remains of Ruscinian mammals (the Spilia locality).

SPECIES DESCRIPTION

Described below are new cardiid taxa and species typical only of the Messinian deposits and never mentioned previously in the Russian publications (Plate).

Family: Cardiidae.

Genus: *Cerastoderma* Poli, 1795.

Cerastoderma macedonicum Nevesskaja, sp. nov.

Plate, nos. 1 and 2

The species is named after Macedonia.

Holotype: PIN no. 4743/1; impression of the left-hand valve on the rock; Macedonia, the Serres basin (Fig. 2; level 1.9); upper Miocene, the Choumnicon Formation.

Description. The shell is 12.5 mm long, rounded (H/l about 0.90), convex, moderately scalene (lap/l slightly more than 0.40), and having the moderately protruding nonkeeled apex. The apical angle is 115°–120°. The total number of ribs is 31–34. The anterior area is decorated with 24–26 rounded moderately convex and closely spaced ribs divided by the intercostal furrows, which are narrower than the ribs. Seven to eight thin and slightly protruding riblets are located in the posterior area. The keel bend is rounded. In the hinge of the left-hand valve, there are two cardinal teeth, but only the frontal one is well developed. The lateral teeth are isolated and platy; one is located behind the nymph far from the apex. The hinge of the right-hand valve and the character of the inner surface are unknown. The shell has a third (outer) layer typical of cerastoderms and distinguishing them from the brackish-water cardiids.

Comparison and remarks. The species differs from *Cerastoderma edule* (Linnaeus) and *C. glaucum* (Poiret) by a greater number of ribs (the total number of ribs on species under comparison is 17–30, and less than 20 in the anterior area) and by the absence of obvious squamose sculpture. Similar forms from the Badenian deposits appear to be determined as *"Cardium" ruthenicum* Hilber.

Distribution: the upper Miocene of Greece.

Material: impressions of two valves.

Lymnocardium (Euxinicardium) prominens Nevesskaja, sp. nov.

Plate, nos. 3–5.

The species name means prominent (Latin, prominensis).

Holotype: PIN no. 4743/10, the left-hand valve impression; Macedonia, the Serres basin (Fig. 2, level 1.15); upper Miocene, the Choumnicon Formation.

Description. The shell is up to 14 mm long, short and trapeziform (H/l 0.86–1.02), notably convex, with unequal sides (lap/l 0.26–0.27) and notably prominent apex (h/H 0.15 or slightly larger). The apical angle is 100°–115°. In the anterior area, there are 12–15 high ribs, convex in front and triangular in the keel part. The space between the frontal ribs is filiform and almost as wide as the ribs in the keel part. The anterior area is decorated with 4–5 tiny riblets. The hinge is unknown.

Comparison. The species is most similar to *L. (E.) seminski* (Andrus.) Ebers., *L. (E.) pseudosecans* Andrus, and *L. (E.) kalidjanicum* Andrus. It differs from the first two species by its shorter shell and from *L. (E.) seninski* also by the ribs less widening downwards. From *L. (E.) kalidjanicum* the species differs by a larger number of ribs (the former has eight to nine ribs in the anterior area).

Distribution: the upper Miocene of Greece.

Material: three valve impressions

Lymnocardium (Euxinicardium) serresense Nevesskaja, sp. nov.

Plate, nos. 6–7.

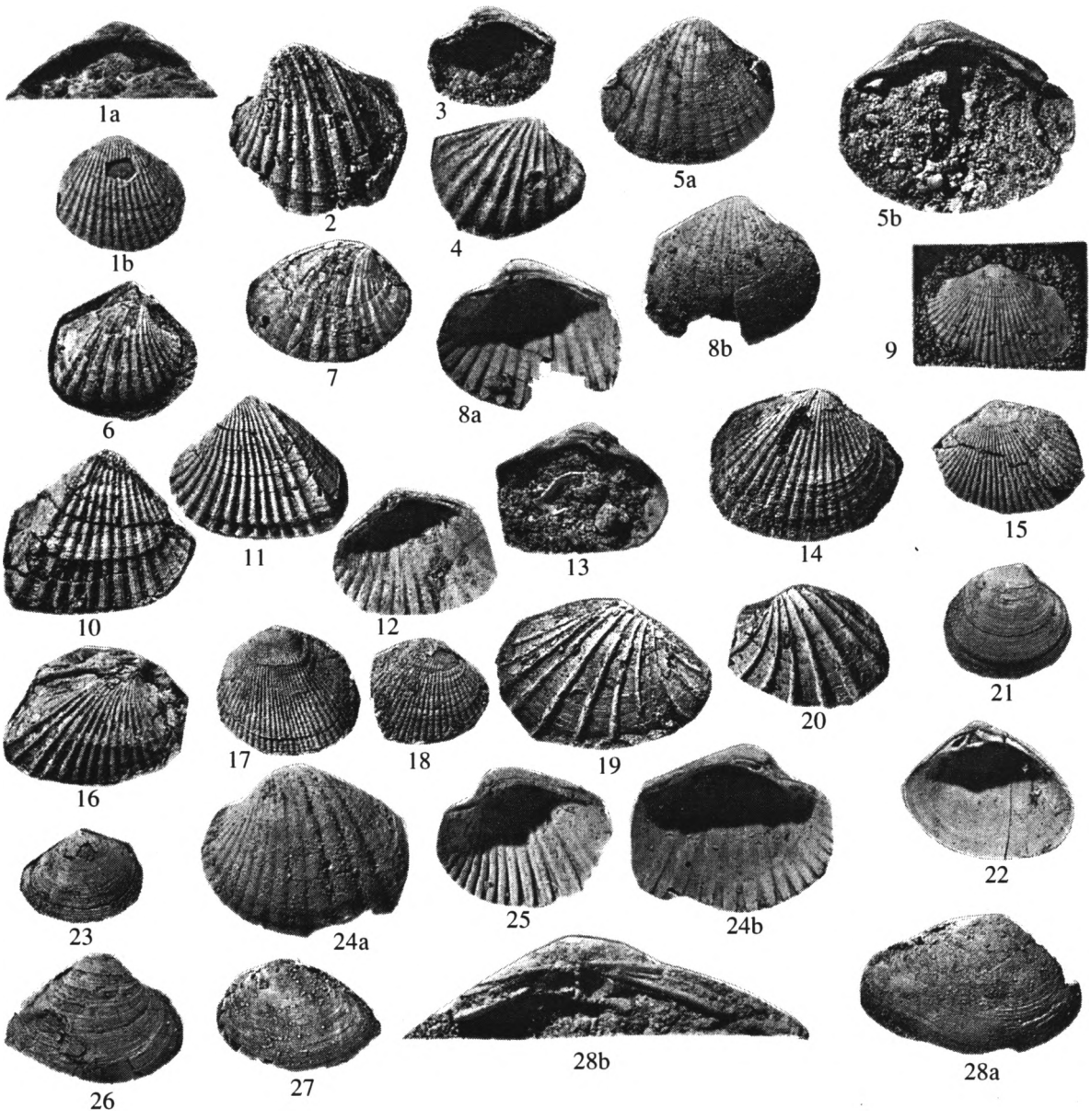
The species is named after the Serres locality.

?Lymnocardium sp. nov. 2; (Gillet and Geissert, 1971, p.142; Table 13; Fig. 2).

Holotype: PIN no. 4743/13, the left-hand valve impression, Macedonia, the Serres basin (Fig. 2; level 1.12); upper Miocene, the Choumnicon Formation.

Description. The shell is up to 10.5 mm long, trapeziform (H/l 0.77–0.88), moderately convex and scalene; its apex is moderately prominent. The apical angle is 110°–115°. The total number of ribs is 20–25. In the anterior area, there are 14–20 flattened-convex ribs; ribs at the keel bend are slightly asymmetrical in cross section. The intercostal space is somewhat narrower than the ribs. In the posterior area, there are five to six slightly prominent riblets. The hinge is unknown.

Comparison. The species differs from *L. (E.) nobile* Stef. and *L. (E.) inlongaevum* Ebers., most closely related to it in shape and character of ribs,



because of the flattened-convex, but not triangular, rib profiles.

Distribution: the upper Miocene of Greece.

Material: four valve impressions.

Subgenus *Hellenicardium* Nevesskaja, subgen. nov.

Typical species: *Lymnocardium (Hellenicardium) hellenicum* Nevesskaja, sp. nov., the Serres basin; upper Miocene.

Description. The shell is rounded-triangular, moderately to notably convex, displaced toward to the anterior edge, and nonkeeled. The outer surface is covered with slightly prominent ribs. The hinge of the right-hand valve consists of separate anterior and posterior lateral teeth, and of one or two small teeth; the anterior tooth is most developed. The hinge of the left-hand valve is unknown. The nymph is narrow and short. The mantle line morphology is unknown.

Genus composition: type species.

Comparison. The taxon differs from the *Ecericardium* subgenus of a similar character of ribbing by the poorly developed hinge.

Distribution: the upper Miocene of Greece.

Lymnocardium (Hellenicardium) hellenicum Nevesskaja sp. nov.

Plate, no. 8.

The species name "hellenic" means Greek in English.

The brackish-water bivalve mollusks from northern Greece: (1) *Cerastoderma macedonicum* Nevesskaja, sp. nov., holotype PIN no. 4743/1, left-hand valve on the rock, Khristos locality (level 1.9), external view, ×2 (a) and hinge view, ×3 (b); (2) *Lymnocardium (Euxinicardium) prominens* Nevesskaja, sp. nov., holotype PIN no. 4743/10, left-hand valve on the rock, ×2, Labirintos locality (1.15); (3 and 4) *Lymnocardium (Euxinicardium) nobile* Stefanescu, (3) no. 4743/9, broken left-hand valve, internal view, ×2. Labirintos locality (1.11), and (4) no. 4743/8, right-hand valve on the rock, ×2, Labirintos locality (1.12); (5) *Lymnocardium (Hellenicardium) hellenicum* Nevesskaja, sp. nov., holotype PIN no. 4743/19, right-hand valve, external view, ×1.5 (a), and internal view, ×2 (b), Labirintos locality (1.13); (6) *Lymnocardium (Euxinicardium) serrense* Nevesskaja, sp. nov., holotype PIN no. 4743/14, right-hand valve on the rock, ×2, Labirintos locality (1.12); (7) *Lymnocardium (?Euxinicardium)* cf. *almeri* Gillet., holotype PIN no. 4743/18, right-hand valve on the rock, ×1, Labirintos locality (1.9); (8) *Lymnocardium (Hellenicardium) astrimonense* Nevesskaja, sp.nov., holotype PIN no. 4643/20, right-hand valve, ×3, internal view (a) and external view (b), Labirintos locality (1.11); (9) *Pseudocatillus subdentatus* (Deshayes), PIN no. 4743/35, left-hand valve on the rock, ×1, Khristos locality (1.10); (10–13) *Pontalmyra incerta chiae* Almera, (10) PIN no. 4743/21, right-hand valve, external view, ×1.5, Labirintos locality (1.14); (11) PIN no. 4743/22, left -hand valve, external view, ×1.5, ditto; (12) PIN no. 4743/24, right-hand valve, internal view, ×2, ditto; (13) PIN no. 4743/25, left-hand valve, internal view, ×2. ditto; (14 and 15) *Pseudocatillus submedius* Nevesskaja, sp. nov., (14) holotype PIN no. 4743/31, right-hand valve on the rock, external view, ×1.5, Labirintos locality (1.8), and (15) PIN no. 4743/32, left-hand valve on the rock, external view, ×1.5, ditto; (16) *Pseudocatillus pseudocatillus* (Barbot de Marny), PIN no. 4743/27, cast of the right-hand valve, ×1.5, Souvalia locality, Aegina I. (2.2); (17 and 18) *Pseudocatillus multicostatus* (Gillet et Geissert), (17) PIN no. 4743/36 left-hand valve, external view, ×1.5, Labirintos locality (1.11), and (18) PIN no. 4743/40, right-hand valve on the rock, external view, ×1.5, Khristos locality (1.9); (19 and 20) *Paradacna abichi* (R. Hoernes), (19) PIN no. 4743/42, right-hand valve on the rock, ×1.5, Labirintos locality (1.12), and (20) PIN no. 4743/46, left-hand valve on the rock, external view of transitional form to *P. huoti* (Andrus.) Nevessk., ×1.5, Labirintos locality (1.12); (21 and 22) *Parvivenus widhalmi* (Sinzov), (21) PIN no. 4743/50, right-hand valve, external view, ×2, Labirintos locality (1.8), and (22) PIN no. 4743/33, right-hand valve, internal view, ×2.5, Labirintos locality (1.9); (23) *Abra* cf. *tellinoides* (Sinzov), PIN no. 4743/ 48, right-hand valve, external view, ×1.5, Khristos locality (1.8); (24 and 25) *Eupatorina littoralis* (Eichwald), (24) PIN no. 4743/4, right-hand valve, ×3, external (a) and internal (b) views, locality Labirintos (1.11), and (25) PIN no. 4743/7, right-hand valve, internal view, ×3, Labirintos locality (1.13); (26–28) *Mactra* cf. *fougeresi* Gillet et Geissert, (26) PIN no. 4743/55. right-hand valve, external view, ×1, Labirintos locality (1.12), (27) PIN no. 4743/59, left-hand valve, external view, ×1, Labirintos locality (1.9), and (28) PIN no. 4743/61 right-hand valve, external view, ×1 (a), and hinge view, ×3 (b), Labirintos locality (1.12).

Holotype: PIN no. 4743/19, the right-hand valve; Macedonia, the Serres basin (Fig. 2, level 1.13); upper Miocene, the Choumnicon Formation.

Description. The shell is 17 mm long, rounded-triangular (H/l 0.87), moderately concave (c/H 0.36), and notably scalene (lap/l 0.34); it has a prominent apex (h/H 0.11). The apical angle is 110°. The total number of ribs is somewhat more than 20; in the anterior area, there are 15 slightly prominent and almost flat ribs; five to six almost imperceptible ribs are noticeable in the posterior area. The intercostal space is narrow. The hinge of the right-hand valve includes the well-developed anterior lateral tooth, the small cardinal tooth, and the posterior lateral tooth. The nymph is very narrow and short. The comparison with other species of the subgenus is presented above.

Distribution: the upper Miocene of Greece.

Material: one valve.

Lymnocardium (Hellenicardium) strimonense Nevesskaja, sp.nov.

Plate, no. 10

The species is named after the Strimon River.

Holotype: PIN no. 4743/20, the right-hand valve; Macedonia, the Serres basin (Fig.2; level 1.11); upper Miocene; the Choumnicon Formation.

Description. The shell is up to 8 mm long, rounded-triangular (H/l 0.77), markedly convex (c/H 0.45), moderately scalene (lap/l 0.40), and having a prominent apex (h/H 0.11). The apical angle is 120°. The total number of ribs is about 20; in the anterior area, there are 16 almost flat or slightly convex ribs, while the ribs of the posterior area are completely flat. The intercostal space is narrower than the ribs. The hinge of the right-hand valve has the anterior and posterior lateral teeth and two small cardinal teeth. The nymph is narrow and short.

Comparison. The species differs from the type species by a longer and less scalene shell having two cardinal teeth in the right-hand valve.

Distribution: the upper Miocene of Greece.

Material: one valve.

Pontalmyra incerta chiae (Almera, 1894).

Plate, nos. 11–14.

Didacna chiae (Gillet, 1965; p. 32, Table 2, Figs. 11, 12; Table 5, Fig. 1–13; Table 7, Figs. 1, 2, 17).

Limnocardium (s.l.) *carinatum* (Cavello et Repetto, 1992; p. 207, Fig. 612).

Description. Shell are up to 25-mm long, angular to trapeziform-oval or rounded-triangular (H/l 0.70–0.95), moderately or markedly convex (c/H 0.30–0.45), moderately scalene, (lap/l 0.35–046), with an average or prominent keeled apex (h/H 0.07–0.13). The apical angle is 100°–130°. The total number of ribs is 19–29. In the anterior area, they have 14–22 triangular or rounded-triangular ribs; the keel rib is prominent and decorated with spinules. The intercostal spaces are flat and usually narrower than the ribs. In the posterior area, there are five to eight prominent ribs; the last posterior rib is sometimes covered with spinules. The keel is usually sharp all along its length. The hinge of the right-hand valve consists of one cardinal and thin solitary anterior and posterior lateral teeth. The left-hand valve

has one cardinal tooth. The nymph is thin, extending for about 1/3 of the length of the posterior branch of the hinge edge. The mantle line is entire.

Comparison. The form is closest to *P. incerta fragilis* (Andrus.), from which it differs by a shorter length.

Comments. The *chiae* and *fragilis* subspecies are assumed to be synonymous, but owing to the remoteness of their distribution areas and different stratigraphic ranges, this conclusion can only be accepted with caution.

Distribution: the upper Miocene or the Messinian of Spain, Italy, and Greece.

Material: about a hundred shell impressions and several separate valves.

Pseudocatillus submedius Nevesskaja, sp. nov.

Plate, nos. 15–16.

The species name is a modification of *P. medius*, in which prefix sub means almost in Latin).

Holotype: PIN no. 4743/31, the right-hand valve; Macedonia, the Strimon basin (Fig. 2, level 1.8); upper Miocene, the Choumnicon Formation.

Description. The shell is up to 17.5 mm long, rounded-trapeziform or angular-oval (H/l 0.82–0.90), moderately convex and scalene (lap/l 0.35–0.40); its apex is slightly to moderately prominent. The apical angle is 130°–135°. The total number of ribs is 31–38. In the anterior area, there are 23–30 flattened ribs divided by the filiform intercostal depressions. In the posterior area, there are 8–10 slightly prominent ribs. The keel bend is obscure. The hinge of the right-hand valve has one small cardinal tooth and thin lateral anterior and posterior teeth; the left-hand valve has one underdeveloped cardinal tooth. The nymph is thin and short. The character of the mantle line is unknown.

Comparison. The species is distinguished from *P. medius* Ebersin owing to a shorter shell (*P. medius* is characterized by H/l 0.62–0.72) and a lesser number of ribs (as compared to 32–45 ribs typical of the latter). It also differs from *P. pseudocatillus* by its considerably shorter and less scalene shell.

Distribution: the upper Miocene of Greece.

Material: about 10 shells on the rock.

Pseudocatillus multicostatus (Gillet et Geissert, 1971).

Plate, nos. 17–19.

Pontalmyra multicostata (Gillet and Geissert, 1971; p. 149, Table 15, Figs. 13 and 14; Table 16, Fig. 1; Table 19, Fig. 6).

Description. The shell is up to 16 mm long, rounded, less frequently oval, short or moderately elongated (H/l 0.74–0.96) and convex, but distinctly scalene (lap/l 0.38–0.42); it has the moderately protruding nonkeeled apex (h/H about 0.10). The apical angle is 115°–130°. The total number of thin closely spaced ribs is 40–50. The intercostal depressions are filiform. The keel bend is perceptible only at the apex. The hinge is poorly developed and is composed of only one small cardinal tooth in each valve. The nymph is very thin. The mantle line shape is unknown.

Comparison. By the character of ribbing and hinge, the species is most similar to *P. polemonis* (Schwets), but the latter is considerably longer and more scalene.

Distribution: the upper Miocene of Greece.

Material: about 20 shells on the rock.

Mactra cf. *faugeresi* Gillet et Geissert, 1971.

Plate, nos. 22–23.

Mactra faugeresi (Gillet et Geissert, 1971; p. 155; Table 18[13]; Figs. 2–9).

Description. The shell is up to 38 mm long, triangular-oval, and markedly elongated (H/l 0.61–0.79); in the larger specimens, the posterior part of the shell is narrowing and turning upwards (beaklike); this more or less convex part is either notably or moderately scalene (lap/l 0.31–0.45). The apex is slightly prominent; the keel bend is angular, and the area behind the keel is gently sloping and slightly concave. The right-hand valve has the paired lateral and two cardinal teeth. The hinge morphology of the left-hand valve and the inner valve surface are unknown.

Comparison. From the Sarmatian *Mactra caspia* similar in morphology, the species differs by the posterior edge pulled back.

Distribution: the upper Miocene of Greece.

Material: about 20 valves on the rock.

DISCUSSION

The obtained results are important for discussion of paleogeographic connections between the Tethys and Paratethys. They also elucidate the origin of the Pontian and late Messinian brackish-water faunas, and the Pontian position in the general stratigraphic scale.

The earlier idea (Stevanovic and Il'ina, 1982; Nevesskaya *et al.*, 1986; Kojumdgieva, 1987; *Pontien...*, 1989) suggesting existence of the Thracian-Aegean Gulf in the Eastern Paratethys at the end of the Tortonian-Messinian, or at the beginning of the Pliocene (during the Maeotian-Pontian, or even in the Sarmatian) appears to be not quite plausible. Although the fauna interchange with the Paratethys could have existed through this gulf both in the early Maeotian and during the Pontian, the Aegean basin had closer connections with the Mediterranean. This is evident from the normal pelagic euryhaline fauna typical of beds underlying and overlying the brackish-water deposits, and also from a considerable difference in the composition of their brackish-water mollusks and molluscan fauna characteristic of the Pontian. *Cerastoderma* and *Mactra* genera that dwelled in the Aegean basin

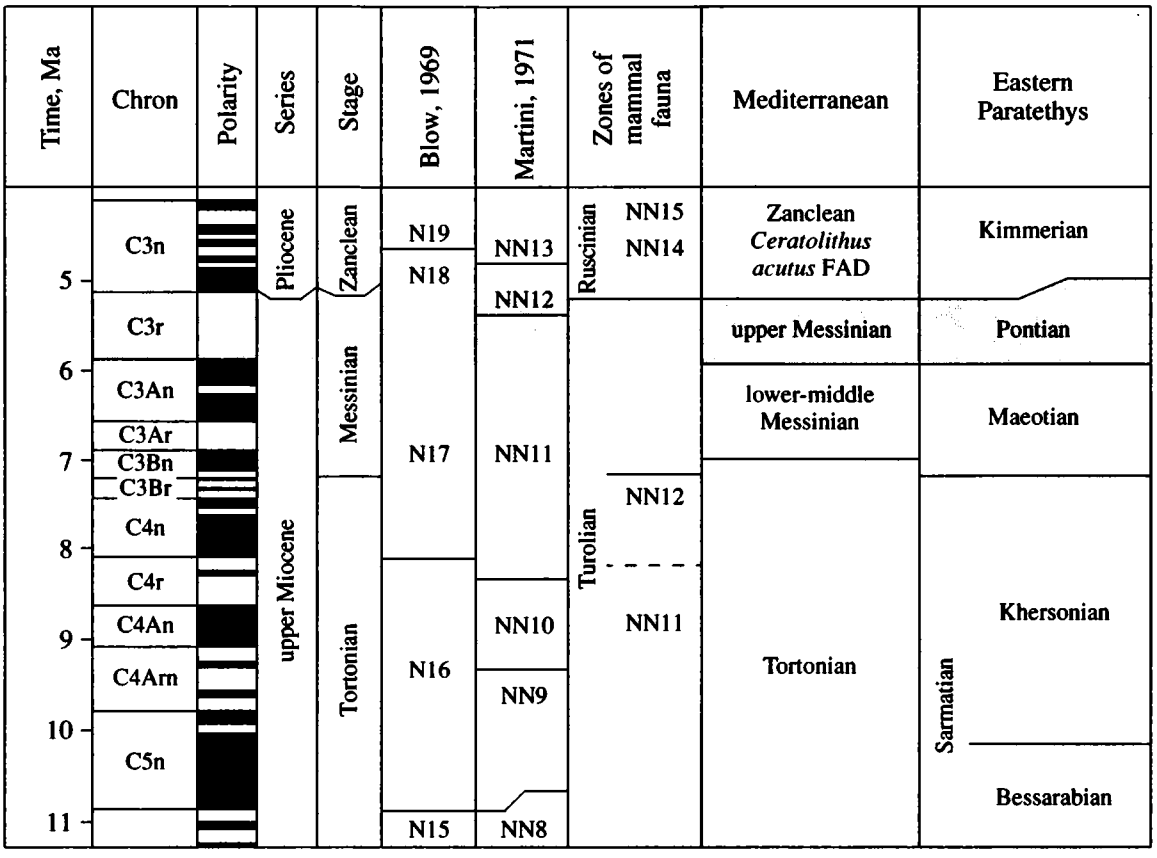


Fig. 3. Correlated upper Miocene stratigraphy in the Mediterranean and Eastern Paratethys (data sources: Berggren *et al.*, 1995; Esu and Taviani, 1989; *Pontien...*, 1989).

together with species unknown in the Paratethys, but typical of the Messinian in Spain and Italy, indicate that the arrival of this fauna from the Paratethys is impossible. Apparently, the Aegean basin served as an intermediate link between the Mediterranean basin of the Messinian time, in which (or in the northern periphery of which) this fauna first originated, and the Euxinian-Caspian basin of the Pontian time, where it developed further. The Aegean basin could be also an intermediate habitat area for the fauna of the early Maeotian Paratethys. Probably, there were several migration waves of this fauna to the Eastern Paratethys. The larger part of brackish-water mollusks in common with others appeared in the Paratethys since the beginning of the Pontian, but *Lymnocardium (Euxinicardium) nobile* and *Pseudocatillus subdentatus* are known only from the upper Pontian (Portapherian and Bosphorian) beds. The fauna migrations in the Portapherian time were likewise implied by data on ostracodes (data of M. Stancheva in Roegl *et al.*, 1991).

Time ranges of the Pontian and Maeotian, like their correlation with the Mediterranean stratigraphic scale, are the most acute topics in discussions about the Paratethys stratigraphy. Due to the reversed magnetic polarity in the larger part of the Pontian interval in the Eastern Paratethys, it was correlated by paleomagnetic data either with the Chron 6 of the Harland's scale, and thus with the upper Tortonian and lower Messinian (according to M.A. Pevzner), or with the lowermost part of chron 4 (C3r) and, consequently, with the upper Messinian (according data of V.M. Trubikhin from *Pontien...*, 1989). The point of view approaching the former one was adopted in a volume of *Stratigraphy of the USSR* (*Neogenovaya...*, 1986) and supported by radiometric and fission track dating (Chumakov, 1993).

In addition to the general identity of specific brackish-water ostracodes shown in the earlier publications (Gramann and Kockel, 1969; Krstic and Stancheva in *Pontien...*, 1989; Roegl *et al.*, 1991), a certain similarity in generic composition of brackish-water mollusks and their species in common indubitably prove that the Choumnicon Formation and deposits studied in the Aegina Island are coeval with sediments accumulated in the Paratethys during the Pontian (or during a part of this stage). On the other hand, the studied assemblages are similarly close to the "Pontian" faunas of the southeastern Spain (Almera, 1889; Gillet, 1965), and their host beds may be coeval with the upper Messinian *Congeria* Beds in Piedmont (Cavallo and Repetto, 1992) and Sicily (Geronimo *et al.*, 1991), and also with the upper Messinian Columbachi Formation of northern Alps (Esu and Taviani, 1989). We should acknowledge,

however, that it is extremely complicated to discuss the similarity of all these assemblages at the species level, owing to the different preservation degree of paleontological material, the quality of published images, and a great variability of species. A confident conclusion is attainable here only after the purposeful study of the fossils themselves. Nevertheless, the generic similarity and presence of specific brackish-water cardiids in all these assemblages, several species of which were common in the Pontian Paratethys and Aegina basin, can be acknowledged now without doubts. Thus, we are inclined, on the basis of presented data, to adopt the correlation of the Pontian with the upper Messinian *Congeria* Beds (Fig. 3), as it was suggested by Esu and Taviani (1989), and also by P.M. Stevanovic and V.M. Trubikhin (see in *Pontien...*, 1989; p. 351).

CONCLUSION

(1) In the Serres basin of Macedonia, 22 species of bivalve mollusks and one gastropod species were identified as characteristic of brackish-water deposits of the Choumnicon Formation. Six species and one subgenus of cardiids are described as new (see the table). In the Souvalia section of the Aegina Island, eight bivalve species were determined.

(2) In the Serres basin, the brackish-water sediments about 40 m thick overlie the Dafni Formation of marine deposits with euryhaline fauna. In both studied regions, these sediments are overlaid by deposits also bearing the marine fauna and plankton; nannoplankton from the latter includes *Ceratolithus acutus* discovered in Macedonia (near Saloniki) and the Aegina Island (Steffens *et al.*, 1979; Roegl *et al.* 1991).

(3) As evidenced by the composition of specific brackish-water mollusks, the cardiids first of all, the concurrent accumulation of all these deposits in the Aegean basin is doubtless, and this is also correct with respect to deposits of the Trilophos site near Saloniki (Gillet and Gessert, 1971). The similarity in generic composition of molluscan assemblages and their species in common allow us to correlate the enclosing rocks with the Pontian strata of the Eastern Paratethys (or with their part), on the one hand, and with the upper brackish-water part of the Messinian sections, on the other.

(4) The euryhaline marine facies, representatives of *Cerastoderma* and *Mactra* genera in the brackish-water molluscan assemblages, benthonic foraminifers untypical of the Pontian, and considerable differences in the species composition discredit the idea suggesting the former existence of the Thrakian-Aegean gulf in the Paratethys. The Aegean basin was more closely connected with the Mediterranean (Messinian) basin.

(5) In all probability, it was the Aegean Sea that served as an intermediate basin, from which the impoverished marine fauna typical of the early Maeotian migrated together with the Pontian fauna into the Eastern Paratethys, where the latter developed further.

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Changes in the Organic World and Geomagnetic Field during the Vendian–Phanerozoic

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Attempts to relate processes occurring at the Earth surface, in particular, changes in the biosphere with the magnetic field behavior, i.e., with processes in the core and deep mantle of the Earth, have been going on for the past several decades. Works of this kind are oriented in two directions. (1) Some scientists tried to find the direct correlation of biostratigraphic levels with the magnetic field reversals considering this as a straight influence of significant changes in the geomagnetic field on the organic world. The exact coincidence of sudden changes in biota (extinction or transformation of old species and/or appearance of new forms, etc.) with the field reversals was established only in rare cases (e.g. Khramov *et al.*, 1982; Merrill and McElhinny, 1983), and consequently, such a simplified understanding of important changes in organic world as "straightly" dependent on the magnetic field reversals *only* is hardly acceptable. (2) Tectonic, paleogeographic, climatic, biological, and other secular changes, i.e., the events and phenomena in the lithosphere and at the Earth surface were compared in other works with the behavior of a geomagnetic field reflecting specific processes in the core and lower mantle (Khramov *et al.*, 1982; Merrill and McElhinny,1983; Rampino and Caldeira, 1993; Pecherskii and Didenko, 1995; and others). Moreover, many data indicate that movements at the core/mantle level and in lithosphere domains are almost synchronous. For instance, some specific events in the plate movements, particular stages of major biospheric events, long-term eustatic sea-level fluctuations, changes in spreading rates and reversal frequency occur within relatively narrow time intervals; the time-shift here does not exceed 10 m.y. (Vogt, 1972, 1975; Aparin, 1982; Danukalov *et al.*, 1983; McFadden and Merrill, 1986; Gaffin, 1987; Pecherskii and Nechaeva, 1988; Loper *et al.*, 1988; Pecherskii and Didenko, 1995; Ricou and Gibert, 1997; and others). However, other scientists registered a significant delay for 10–60 m.y. between the commencement of geological eras and acceleration in geomagnetic reversals frequency, which was noticed for the Phanerozoic (Molostovskii *et al.*, 1976; Khramov *et al.*, 1982), Paleozoic (Didenko, 1998), and entire Neogean (Pecherskii, 1999). Synchronism and delays characterizing these phenomena were discussed by researchers during the last 20 years. This period resulted in a noticeable modification of the chronostratigraphic scale, where the boundaries of stage, system, and era changed their positions and dates, but despite this, the aforementioned regularities continued to be registered, thus suggesting that both the synchronism and time gaps between some processes and events occurring at the surface of the Earth and near its core really do exist. Accordingly, two principally different types of mechanisms are responsible for the processes occurring on and within the Earth (Pecherskii, 1999). Mechanisms of the first type act synchronously at the surface and near the core of the Earth; this can be manifested, for instance, in variations of the rotation velocity and inclination of the Earth's axis (Pecherskii and Nechaeva, 1988; Pecherskii and Didenko, 1995). Mechanisms of the second type function at the core/mantle boundary, and their "echo" reach the Earth's surface with the velocity comparable with that characteristic of the continent movement (Khramov *et al.*, 1982; Merrill and McElhinny,1983; Loper *et al.*, 1988; Rampino and Caldeira, 1993; and others).

Comparative rates of changes in the organic world and geomagnetic field polarity during the Vendian–Phanerozoic. One way or the other, the above-mentioned geological processes are reflected in development of the organic world and, thus, on a chronostratigraphic scale. Let us try to confirm (or discredit) the above-mentioned relationship between processes occurring at the surface of the Earth and near its core by comparing general rates of the organic world evolution and geomagnetic field variations during the Phanerozoic and Vendian. This can be done using two main data sources. One of them is a summary by Benton (1995) that provides data on rates (gradients) of changes in the diversification and extinction of organisms (a number of families per every 10 m. y.). For our purposes, most useful are data on marine organisms, since they most completely characterize the Vendian and entire Phanerozoic. The second source is the geochronological time scale (Harland *et al.*, 1990) that is

used here with some modifications for the Vendian-Cambrian interval to calculate the number of stages per every 10 m. y. According to data by Grotzinger *et al.* (1995), Kaufmann and Knoll (1995), and Rozanov *et al.* (1997), the following dates for lower boundaries of the Vendian-Cambrian stages are accepted: 610 Ma for the Lapplandian (Varanger), 570 Ma for the Redkino Horizon, 551 Ma for the Kotlino Horizon, 544 Ma for the Rovno (Nemakit-Daldynian) Horizon, 530 Ma for the Tommotian, 528 Ma for the Atdabanian, 525 Ma for the Botomian, 520 Ma for the Middle Cambrian (divided into two stages in the middle), 510 Ma for the Upper Cambrian (divided into two stages in the middle), and 505 Ma for the Tremadocian. Of course, the accepted scheme is not indisputable, but the accuracy of every boundary position is not essential for our analysis; more important is a comparison of biostratigraphic and geomagnetic information calibrated relative to the common time scale. Correspondingly, the integral scale of geomagnetic polarity that was proposed earlier (Pecherskii, 1997) is modified in line with the above-mentioned Vendian-Cambrian chronostratigraphic scale (Pecherskii, 1999). In addition, the scale takes into consideration magnetostratigraphic data on some sections of Siberia (Kirschvink and Rozanov, 1984; Galle and Pavlov, 1999), the South Urals and Dniester regions (Tretyak *et al.*, 1996). In the last region, data characterize the Vendian stratotype section, but unfortunately, their methodical substantiation does not completely conform to modern requirements, although general patterns are similar to those established in the South Urals (Danukalov *et al.*, 1983). The geomagnetic polarity scale within the mentioned intervals is slightly modified at present (Pecherskii, 1999); however, changes in the frequency of geomagnetic reversals remained almost the same. This indicates that geomagnetic field behavior, when averaged for intervals of 10 m.y. and more, reflects adequately well reality.

To receive a more objective picture, the relative changes of the magnetic polarity sign were depicted using another method (McElhinny, 1971), when the ratios between the number of individual paleomagnetic determinations (collections), including both the field polarities and their total number in the given time interval (in our case, 10 m.y.), were calculated. This method requires an extensive database, but even in this case, the result is relative, qualitatively (not quantitatively) characterizing the frequency of changes in the magnetic polarity sign in terms "many-few", or, to be more exact, "frequent-rare"; correspondingly, only time-positions of extremums, but not their amplitude, can be considered as representative. For this analysis, I used the Global paleomagnetic database (McElhinny and Lock, 1991, 1993) and then compared the obtained result with the curve plotted using the same method for the most part of the Phanerozoic (Johnson *et al.*, 1995). Despite the difference in approach to data selection, both curves virtually coincide one with another.

A comparison of curves, demonstrating the frequency behavior of magnetic field reversals and plotted on the basis of two virtually autonomous methods, shows that the time positions of corresponding extremums (maximums or minimums) in the Vendian and Phanerozoic are almost completely synchronized, but they essentially differ in the amplitude: even in the most reliable Late Paleozoic–Cenozoic portion of the geomagnetic polarity scale, the amplitude of variations in "the frequency of the magnetic field reversals" determined by the second method is several times higher as compared with that obtained using the geomagnetic polarity scale (Fig. 1a, b). Thus, both methods confirm the objectivity of the established rhythms of geomagnetic reversals during the Vendian and Phanerozoic, which are manifested in the alternation of maximums and minimums; the quantitative evaluation of the amplitude of the field reversal frequency is possible only when using the curve plotted with the help of the first method, i.e., on the basis of the geomagnetic polarity scale (Fig. 1a).

As was noted previously (Pecherskii, 1998), the frequency of changes in the geomagnetic field polarity, preponderance of one sign, and relative changes in the intensity value, distinctly correlate with each other. Consequently, there is a synchronism between the *major long-period changes in the value and direction of the geomagnetic field; i.e., these changes are characteristic of the complete field vector.*

Further, let us compare the behavior of a geomagnetic field with changes in the organic world using the above-mentioned characteristics: (1) changes in the number of stages in the chronostratigraphic scale (Fig. 1c); (2) changes in the number of families of marine organisms (Fig. 1d), and (3) extinctions of families of marine organisms in every 10 m.y. (Fig. 1e). To eliminate possible errors in data and age determinations, the smoothing interval was accepted as being equal to 30 m. y. (Fig. 1).

The almost complete coincidence, i.e., synchronism of maximums and minimums in rates of the organic world changes with the magnetic reversals frequency during the Vendian–Phanerozoic, should be noted first of all. For instance, the difference between maximums (minimums) in the frequency of field reversals and the closest maximum (or minimum) stage number or diversification and extinction events of organisms is 3.1 ± 9.6, 3.5 ± 6.3, and 1.4 ± 9.5 m. y., respectively. As is seen, the error does not exceed the data grouping interval (10 m. y.). In addition, many boundaries of geological periods (in the accepted geological time scale) span narrow time ranges not more than 10 m. y. long and correspond to episodes of insignificant changes in the frequency of field reversals (Fig. 1a), which are smoothed within average intervals 30 m. y long. Some difference in the dynamics of the organic world changes and magnetic reversals frequency is related, in my opinion, to the peculiar (variable) evolutionary rate of biota diver-

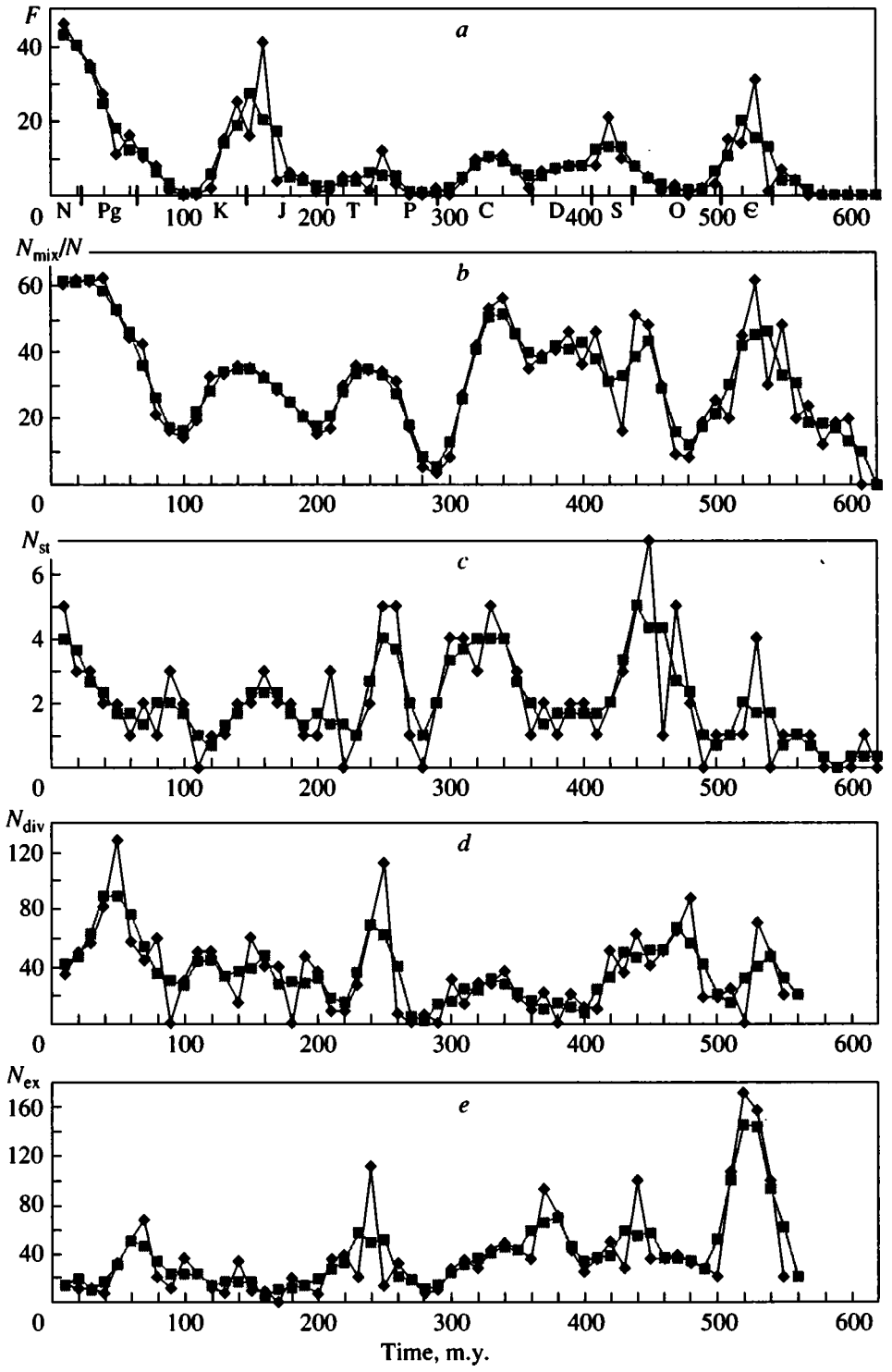

Fig. 1. Dynamics of changes in the organic world and geomagnetic field during the Vendian and Phanerozoic: (a) the number of geomagnetic field reversals in every 10 m.y. (F) calculated using the geomagnetic polarity scale (Pecherskii, 1997, 1999); (b) the rate of relative changes (%) of the geomagnetic field polarity in every 10 m.y. (N_{mix}/N) calculated using the global paleomagnetic data base (McElhinny and Lock, 1991, 1993); (c) the number of stages of the chronostratigraphic scale within every 10 m.y. (N_{st}); (d) changes in the number (diversity) of families of marine organisms in every 10 m.y. (N_{div}); (e) the number of extinct families of marine organisms in every 10 m.y. (N_{ex}). Data are grouped for the intervals 10 m.y. long (diamonds) and smoothed afterward (squares); the averaging interval is 30 m.y.; the step is 10 m. y.

sification and extinction, as well as to the incompleteness of magnetostratigraphic records and erroneous dating results. Thus, it can be stated rather confidently that the acceleration or deceleration of geodynamic processes at the core-mantle boundary are synchronous with remarkable changes of the organic world. Boundaries between geological periods most commonly coincide with periods of decrease in the reversals frequency and in the value of geomagnetic field intensity (but not with particular reversals!). They themselves often mark the rate decrease in the organic world changes (Fig. 1).

If major long-period changes in the geomagnetic field directly influence changes in the organic world, then there should be the noticeable correlation (positive or negative) between magnitudes of synchronous extremums of each parameter characterizing changes in the organic world development and in the magnetic reversal frequency. However, as is seen in Fig. 2, such a correlation between the given parameters or between all of them is virtually absent. Consequently, the long-period changes in the geomagnetic field either do not influence the organic world evolution, or their effect is insignificant.

The idea that the commencement of geological eras is delayed relative to the minimums in the reversal frequency has been suggested previously (Molostovskii *et al.*, 1976; Khramov *et al.*, 1982; Didenko, 1998; Pecherskii, 1999) and is confirmed in this work. As one can see from Fig. 1a, the delay time amounts to 35–60 m. y. and thus corresponds to the rate of the energy transfer from the core-mantle boundary to the Earth surface (approximately 4–10 cm/year). This rate value agrees with the estimate movement velocity of the main continental plates during the Phanerozoic (Zonenshain *et al.*, 1987; Jurdy *et al.*, 1995).

Conclusion. The comparative dynamics of the organic world evolution and geomagnetic field changes during the Vendian–Phanerozoic shows that many geological processes comparable in duration with geological periods and occurring both near the core and at the Earth surface appear to be synchronized. However, it is unlikely that the geomagnetic field variations directly influenced the organic world evolution, or speaking more carefully, were obviously unimportant in this respect. In other words, the considered processes "respond" to the same "external" factor that changes, for instance, the axis inclination and rotation velocity of the Earth. This, in turn, influences processes in the core and lower mantle, i.e., the geomagnetic field behavior, as well as tectonic, climatic, and other events at the Earth surface directly controlling the state of the biosphere. Processes comparable in duration to the geological eras are related to the "internal" mechanism of the energy transfer from the lower part of the mantle upward.

Thus, the performed analysis confirms the idea that processes occurring at the Earth surface and near the core/mantle boundary are closely interrelated. However, this is a result of the action of some common mechanism, and the established connection is acasual.

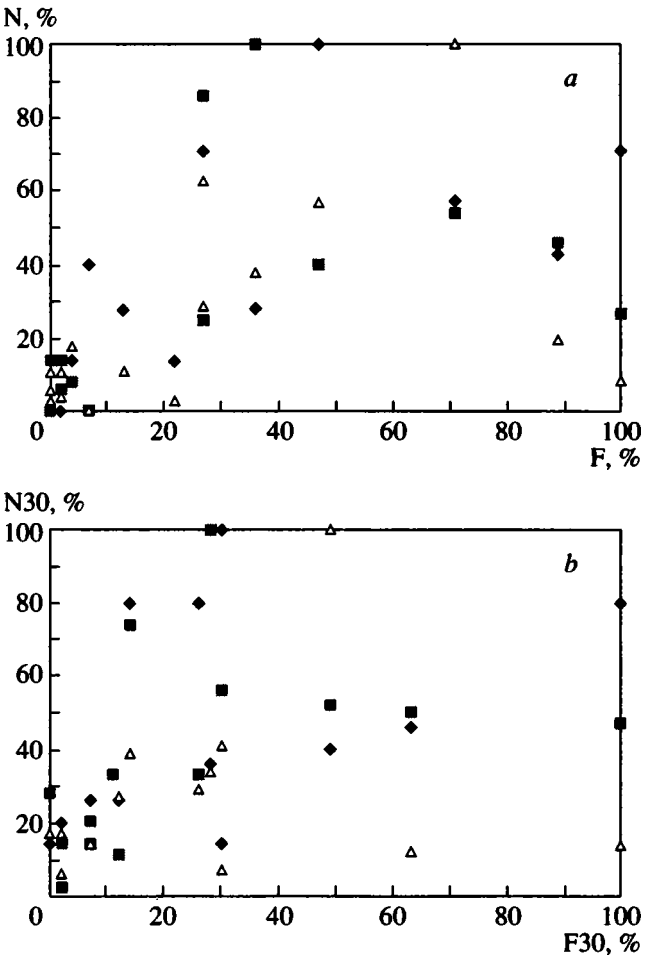

Fig. 2. Correlation diagram showing magnitudes of synchronous extremums of the stage number (diamonds), marine organisms diversity (squares), and extinct families of marine organisms (triangles) versus the extremums of the geomagnetic reversals frequency. For the sake of clearness, ratios between the every parameter and its maximum value are given in %: (a) extremums of the curve characterizing data grouped for the time intervals of 10 m.y.; (b) extremums of the smoothed curve (the averaging interval is 30 m.y.).

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